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SHEARING PHENOMENA AT HIGH PRESSURE OF POSSIBLE IMPORTANCE FOR GEOLOGY

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ABSTRACT

Various phenomena of plastic flow under shearing stress are studied at confining pressures up to 50,000 kg./cm.^{2} The shearing stress required to produce flow may increase many fold beyond its normal value. There may be either "hardening" or "softening" phenomena while a steady state is being reached. In addition to the phenomena of flow, and often simultaneous with them, rupture in shear, followed by self-healing, frequently occurs. This rupture in shear may become more or less pronounced at high pressure. There are various time effects: some substances show very little variation in shearing force for a wide variation of speed of plastic flow, and others show very considerable variations. Self-welding is of almost universal occurrence. Permanent alterations of density are produced, with irreversible transitions to other modifications. Finally, there are various chemical effects of decomposition or synthesis.

Mere inspection of many geological specimens which have come from depth in the earth is sufficient to disclose the universality of flow under the conditions there prevailing, and hence to suggest the importance of a study of the effects of shearing stress, which is necessary to produce flow, under the same conditions. These conditions are high temperature and high pressure in combination. Such combinations are at present difficult of attainment in the laboratory, and studies of various physical phenomena under such conditions have not yet been extended to include the effects of shearing stresses. I have recently been able to study at room temperature various effects of shearing stress at pressures up to 50,000 kg./cm.^{2}, which is the pressure prevailing at a depth of 166 kilometers, assuming an average density of 3. The phenomena found at this temperature

have proved to be both varied and unexpected. Since phenomena at higher temperatures must be at least qualitatively similar in many respects to those at room temperature, we may expect these experiments to indicate some of the essentials of the behavior of material under geological conditions. It therefore seems that some of the results may be appropriately summarized in a periodical for geology. A paper describing some of the results first found has already been published in a periodical of physics,[1] and I hope presently to publish another paper giving various details and many results obtained since writing the first paper.

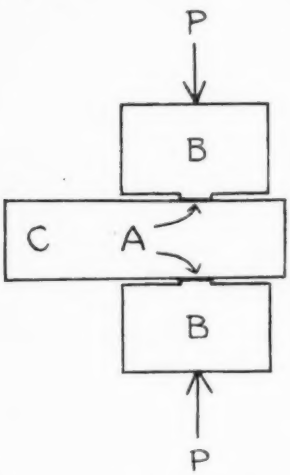

FIG. 1.—Scheme of the apparatus for applying pressure and shearing stress simultaneously.

The essential experimental arrangement is indicated in Figure 1. Two cylindrical blocks, *B* (hereafter called the "pistons"), of hardened steel of the highest physical properties carry very short cylindrical bosses about 6.3 mm. in diameter, ground flat on the face. Between the two cylindrical blocks there is a rectangular block, *C* (hereafter called the "anvil"), of steel of the same quality, ground flat and parallel on the two surfaces facing the blocks. The material to be studied is placed at *A* in the form of a thin disk between each piston and the anvil. The whole combination is mounted under the plunger of a hydraulic press by means of which the pistons can be pushed against the anvil with any desired force up to the maximum which the steel will support, which under these conditions is 50,000 kg./cm.^{2} or even more. The material under examination, *A*, flows out laterally between anvil and piston until its thickness is reduced to such a value that the friction on its flat surfaces holds it in equilibrium against the expelling pressure, which acts only on the thickness. A moment's consideration shows that an equilibrium thickness

[1] P. W. Bridgman, *Phys. Rev.*, Vol. XLVIII (1935), 825–47.

must always be reached if the coefficient of friction is finite. The dimensions of the apparatus were such that the film thickness was very seldom reduced below 0.0025 cm., and for most substances the thickness was several times as great as this.

After the pressure has been applied, the anvil may now be twisted between the pistons, thus applying a shearing stress to the material *A*. At low values of the twisting force the material yields elastically; but if the torque is high enough, the anvil rotates with constant velocity. Only the phenomena in the region of continuous rotation have been studied up until now. The torque required to produce rotation is measured by a simple gauge attached to the handle of the rotating wrench, and from this torque the shearing stress applied to the disks may be calculated. The torque or shearing stress may be plotted as a function of the pressure, and in this way information obtained about the dependence of various shearing phenomena on pressure.

At each pressure the torque was taken as the mean of values given by rotating back and forth through an angle of approximately 60°. Special experiments in which the rotation was continuous and through a number of complete revolutions showed no difference as compared with double rotations through 60°. Rotation was usually by hand, at a speed of perhaps 60° in 5 seconds. Most substances show no perceptible dependence of speed on force under these conditions; but there are some substances in which the variation of speed with force is important, and experiments will be described later in which this effect was studied. With speeds of the order just mentioned, it may be shown by a simple calculation that any temperature effects are completely negligible, the temperature increase amounting at most to only a few degrees.

At very low pressures on the pistons rotation of the anvil is permitted by simple slip of the material *A* on the steel, and the turning moment gives at once the coefficient of friction. In the initial range, in which the coefficient of friction is approximately constant, the curve of torque rises linearly with pressure. With most substances a point is soon reached, however, at which the shearing stress applied to the substance by friction becomes equal to the shearing strength of the material itself. When this point is reached, slip at the surface

ceases, and from here on rotation is accompanied by internal plastic flow of the material. This point is manifested on the curves of "normal" substances by a more or less pronounced knee (Fig. 2). The

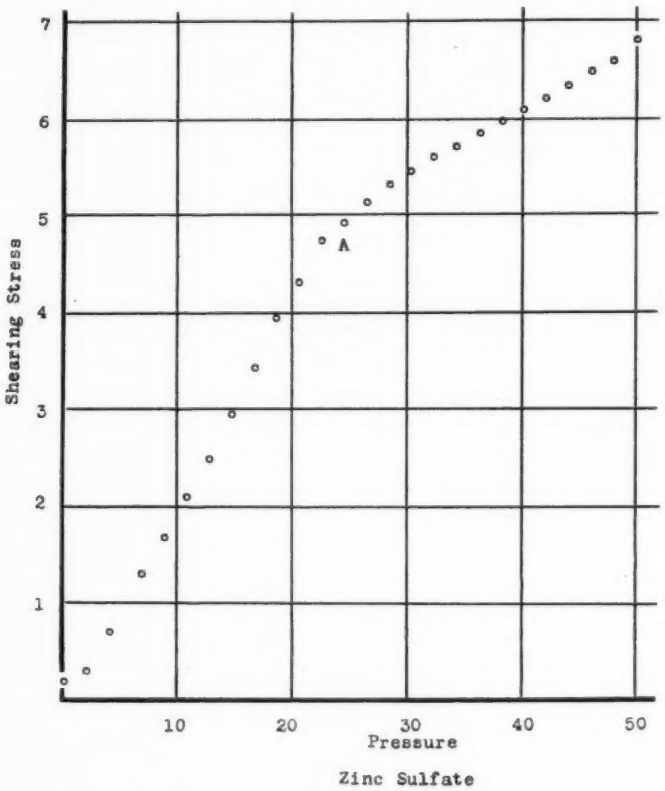

FIG. 2.—Results for zinc sulphate. Ordinates: shearing stress in thousands of kilograms per square centimeter; and abscissas: confining pressure, also in thousands of kilograms per square centimeter.

knee, A, is often rounded instead of sharp, because the pressure is not uniform across the surface of the pistons—all the results refer to rather rough average values for the pressure. Beyond the knee the curve of torque against pressure determines the shearing strength

(i.e., the maximum shearing stress that the material will support without plastic yield) as a function of average pressure.

Engineers have been much concerned, in formulating conditions of rupture, with the shearing stress at plastic flow as a function of pressure, and have come to the conclusion that for many substances it is a sufficiently good approximation for engineering purposes to regard the shearing strength as a constant, independent of pressure. If such were the case under our conditions, the curve of torque against pressure should be a horizontal straight line beyond the knee. This is very far, indeed, from being the case; but the majority of curves have a continuous rise, as indicated in Figure 2. The explanation is merely that the range of stress encountered in ordinary engineering practice is so low compared with that used here that the effect is obscured.

The first fact of geological significance shown by these experiments is just this increase of shearing strength produced by pressure. This increase is many-fold for ordinary substances, and is characteristic of both glasses and crystals. In my paper in the *Physical Review* will be found a table for a number of the metals, showing many cases of an increase of shearing strength by a factor as much as 10 at a pressure of 50,000 kg./cm.^{2} Paraffin, which I had already found, by other means, to have a very great increase of shearing strength at high pressure, is an extreme case of an increase of shearing strength by a factor of at least 900. Since publishing these results for metals, I have measured several hundred salts and a number of minerals; and the same sort of phenomenon is characteristic of them also. Values for shearing strength which rise to figures equal to those for the harder metals are not uncommon. It is not possible in most of these cases, however, to determine the factor by which the strength has increased, because the shearing strength is not known under ordinary conditions.

The fact that shearing strength increases so markedly at high pressures means that it is now possible to study the effects of much higher shearing stresses than were formerly attainable. The maximum shearing stress possible to apply to any substance is, of course, limited to the strength; application of a shearing stress equal to the strength produces plastic flow, and shearing stress higher than this

may not be applied. In this respect the situation is entirely different from that presented by hydrostatic pressure.

The distortion produced in the thin films under these conditions may rise to very high values; thus a film 0.0025 cm. thick rotated through 60° experiences at the outer edge of a 6.3-mm. diameter an angular shearing displacement of 120 radians. It might at first appear that under such extreme conditions the material would become completely amorphous in character, losing all traces of any regular crystal structure; and thus any results obtained might be thought not to apply to ordinary matter but to an effectively new state of matter. This turns out not to be the case, however, for the crystal structure even under these extreme conditions persists through volumes large enough to give the material the properties of ordinary crystalline material. A proof of this is given by the behavior when the material experiences a polymorphic transition. The shearing strength of a substance might be expected in general to change discontinuously when passing through a transition, and such a change would be shown by a break in the curve of torque against pressure. A great many substances have been found to have such breaks, and the fact that the break means a transition has been checked by independent measurement in another apparatus permitting a determination of all the thermodynamic parameters of the transition. The transition, of course, means a change in the lattice structure of the crystal; and such changes could not take place if the material were so severely deformed as to possess no lattice structure.

The large number of polymorphic transitions found at high pressure is of geological significance, at least for that part of the earth's crust in the crystalline condition. Perhaps one-third of the several hundred chemical compounds examined have transitions at room temperature between atmospheric pressure and 50,000 kg./cm.^{2} Furthermore, there is a definite correlation between the melting-point and the number of new forms found, which means that in all probability a number of substances with high melting-points may show transitions at higher temperatures which are not disclosed by the present shearing measurements because internal viscosity at room temperature is so great as to suppress the transition. It is therefore highly probable that under the conditions in the earth

polymorphism is a much more common phenomenon even than yet found in the laboratory.

The volume changes accompanying these transitions may vary from the smallest detectable with the apparatus up to 10, 15, or even 20 per cent. A volume increment or decrement suddenly injected into the earth's crust must produce important disturbances in the neighborhood and in particular zones of shearing stress. This possibly is a mechanism for the production of shearing stress in zones normally under isostatic compensation.

The transitions may be of most various characteristics. The temperature of transition may either increase or decrease with increasing pressure; and I have found a strikingly large number of transitions with temperature nearly independent of pressure, which means a very small latent heat. The velocity of transition may increase or decrease as temperature and pressure change along the transition curve. Transitions which run vertically, that is, with no latent heat, are often characterized by an extremely high, almost explosive, rapidity at the higher temperatures near the melting-point, and a startlingly abrupt sinking-back into somnolence as temperature becomes lower.

There is no rule governing the shearing strengths of the high- and low-pressure modifications; but many examples can be found for either an increase or a decrease of strength on passing from the low-pressure to the high-pressure modification.

There is a theoretical possibility that must be kept in mind. There is no thermodynamic reason why transitions should not occur under shearing stress which would never occur under any combination of temperature and hydrostatic pressure. In the greatly enhanced range of shearing stress realizable at high pressure one may expect greater opportunity for the existence of this phenomenon than in the comparatively low range of shearing stress limited by plastic flow under ordinary conditions. No examples of phenomena which have been positively identified as being of this character have yet been found, but it is not improbable that a very marked break in the shearing curve of lithium may mean a transition of this kind.

In addition to the main phenomena just described, there are a number of secondary phenomena of possibly even greater geological

significance. In the first place, there are transient phenomena while the steady state is being reached. A more or less prolonged seasoning process, consisting of many rotations, is usually necessary before the torque reaches a steady value. During this seasoning process the shearing strength almost always increases. That is, the resistance to plastic flow becomes greater the more finely the structure is broken up by continued deformation. This is just as might be expected. This hardening effect is particularly prominent if the material is originally in the form of a single crystal with the plane of principal slip or cleavage parallel to the face of the piston. For example, if the disk is a thin cleavage flake of zinc or graphite, rotation initially is almost unbelievably easy but hardening rapidly occurs; and in the case of graphite, all the lubricating properties associated with its initially easy internal slip are eventually lost, and at the end graphite becomes an abrasive, actually rubbing metal off the hardened steel surfaces.

After the substance has been completely seasoned by repeated rotations, effects may still persist connected with the direction of rotation. On reversing the direction of rotation after each displacement of 60°, the final value of the rotating force is not always at once attained, but the force may go through a definite pattern of variation. Usually this pattern is quite simple, and consists merely of a slow dropping-down to an asymptotic value from a higher initial value, or a climbing to a higher asymptotic value from a lower initial value. This process is usually completed in the first 20° or less of the rotation. The magnitude is seldom more than a few per cent. Sometimes the sign of the effect reverses with increase of pressure. Sometimes the pattern may be more complicated, such as initial fall and then rise to an asymptotic value lower than the initial value. A single type of detailed explanation cannot be expected to apply to all these cases, but in any event the mere existence of such effects would seem to demand the building-up of some sort of internal structure by the deformation. This again would indicate that the internal condition cannot be entirely amorphous. Even a brief consideration of the geometrical conditions prevailing during plastic flow will suggest that the number and character of the slip planes in the crystal must have an important effect on the nature of the phe-

nomenon. Perhaps in those substances which have a great many slip planes, as cubic crystals, the ultimate steady state will be such that slip is possible on one or another of these planes; and the seasoning process may consist merely of a slight reorientation of these planes in such a way as to be a little more favorable to slip. In such a system, the seasoning process on reversal of direction may be a softening process, slip being easier in the final appropriately oriented condition than initially. On the other hand, if slip is possible on only one plane in the crystal, the geometrical disorientation demanded by the process of plastic flow may ultimately compel yield by tearing across slip planes; and in such a substance the seasoning process on reversing direction of flow might be a hardening process.

Under geological conditions there must often be reversals of direction of shear, and some of these transient phenomena may come into play.

Plastic flow for the majority of substances takes place perfectly quietly and continuously, but in 40 per cent of the cases examined flow is not perfectly smooth. Yield is practically always smooth for cubic crystals, unless the melting-point is very high. This is probably correlated with the large number of possible internal slip planes in the crystal. The most variegated kind of phenomena are shown by those substances which do not flow smoothly. The failure of smoothness may be almost imperceptible, manifested only by a feeling of very fine-grained grinding to the hand on the handle of the rotating wrench; or there may be simply a squeaking during rotation. Or there may be a soft chattering, or a harder and more abruptly punctuated chattering. Or there may be snapping and jumping, sometimes so violent and spasmodic as to make measurements of any quantitative value almost impossible. The phenomena of "snapping" are, I think, of special geological significance. The snapping is merely a manifestation of internal rupture. The material breaks; and the anvil jumps through a finite angle of rotation, the rotating force simultaneously jumping back, because the handle of the rotating wrench or the hand that pushes it are not infinitely stiff. The jump is terminated by the automatic healing of the rupture, and the phenomenon is ready to start over again. This sort of rupture is obviously different in at least one important respect from rupture

under ordinary conditions. Ordinary rupture, as the breaking of a rod in tension, or rupture by twisting the head off a bolt, is accompanied by a change of geometrical configuration, actually division of the specimen; and rupture can occur only once. Under our conditions of shear, however, there is no change of gross geometrical configuration at rupture; the material merely "lets go to get a fresh hold," and the phenomenon repeats itself indefinitely.

These rupture phenomena may or may not be superposed on the ordinary phenomena of plastic flow. Thus there are substances, of which graphite is a conspicuous example, which show no trace of plastic flow; rupture occurs, there is a jump in angular displacement and a jump down in the force, the force then builds up again with practically no further increase in angular displacement until the critical force for rupture is reached, and the process repeats. On the other hand, the majority of substances capable of rupture also have superposed the ordinary phenomena of plastic flow; plastic flow progresses smoothly until a critical value of the angular displacement is reached, when rupture occurs with a snap. Mica is an example of such a substance. During the stage of plastic flow, the force required to maintain rotation may either drop from a higher value to the value at rupture or may climb to the value at rupture from a lower value. The angular displacement during the stage of plastic flow before rupture may vary greatly from substance to substance.

These rupture-and-flow phenomena are not limited by any means to crystalline substances. Thus, SiO_2 glass is a non-crystalline substance which shows only rupture without flow; and on the other hand, borax glass shows a striking combination of both rupture and flow.

It seems to me that the recognition of the possibility of simultaneous existence of two phenomena, rupture and flow, must react on our pictures of how yielding takes place deep in the earth. In particular it would seem that this sort of rupture must be a factor in deep-seated earthquakes. What the precise mechanism is back of the two phenomena cannot be told without more elaborate investigation; but it is not unreasonable to suppose that when fracture occurs in a crystalline material there are different kinds of planes in the crystals, on one of which slip may occur, while on the others only

rupture may occur. Continued plastic flow may result in rotating out of favorable positions those planes on which slip occurs, and rotating into position the rupture planes. Actual rupture throws the system into confusion, and the process of selective orientation starts again. Something qualitatively similar does not seem excluded, even in the case of a glass, without, of course, the degree of regularity characteristic of the crystal.

Slip at the surface between steel and the other substance is accompanied in a number of cases by various sorts of discontinuous phenomena. In such cases rotation at points of the curve below the knee may be accompanied by chattering or creaking, which ceases at once on passing above the knee. More often, however, surface slip is smooth, and the discontinuities occur only above the knee, where there is internal slip.

In the range of internal flow there is often marked variation of the discontinuous effects on increasing pressure. It is more usual for the snapping, etc., to become more violent at higher pressure, but cases occur where violent rupture at a moderate pressure is replaced by smooth plastic flow at high pressure. If the substance passes through a polymorphic transition, effects of this sort are to be expected; and, in fact, several cases have been found of a dramatic cessation of snapping and rupture on passing to the high-pressure modification.

Besides the phenomena of rupture just described, another type of instability is not infrequently encountered. If the initial thickness of the disk between piston and anvil is too great, then, as pressure is increased, the material of the disk may reach a condition of instability and be violently expelled from between anvil and piston. Occasionally such an unstable condition is produced merely by increase of pressure; but more often the maximum pressure may be reached smoothly enough, and then the first touch of rotation is sufficient to introduce enough mobility into the system to induce instability, and the material of the film is expelled with explosive violence. Large amounts of energy may be set free in such mechanical explosions, and secondary temperature effects may be produced sufficient to initiate various chemical effects. Mechanical instability like this can be avoided merely by starting with a thinner disk.

Time effects may play an important part with a number of substances. For most substances in the range of velocities of rotation corresponding to normal patience with manual operation, the torque is independent of speed almost within the limits of error. But there are a few substances for which time effects are prominent in this range of velocities. This led to the construction of special apparatus for studying these effects over a much wider range than possible by ordinary manual operation. Two sorts of investigation were made: One was an investigation at constant torque. Such a torque was produced simply by moving the rotating lever by a weight with cord and pulley, and determining the speed, either at constant weight or as a function of the weight. The second was an investigation at constant speed. The handle was rotated at any desired constant speed by a motor and system of reducing gears, and the force was measured by a strain gauge in the cord attached to the rotating handle. The substances which show the largest time effects are the low-melting metals, such as tin or lead. In Figure 3 is plotted the speed of rotation as a function of torque. Tin is a rather extreme case in which a variation of speed of 10,000 fold is accompanied by a variation of force of 10 fold. The nature of the yield is still essentially that characteristic of a solid, in a liquid the speed being directly proportional to the force; but it obviously does not have much significance to talk about a shearing "strength" or a yield "point" for a substance like tin. If one extrapolates the curve of Figure 3 as one would extrapolate any ordinary curve, one would draw the conclusion that at zero shearing force there would still be a residual rotation, an absurd conclusion. The danger of any extrapolation of laboratory results to geological times is obvious.

The effect of pressure on the rate of yield of tin is also shown in Figure 3. Under a given torque the rate of yield decreases with increasing pressure, as would be expected. There is also a change of shape of the curve at the low torque end, suggesting an increase in the abruptness with which rotation ceases with decrease of force, and therefore a closer approach to what may be characterized as a true strength. Copper, as shown in Figure 3, approaches much more closely to the ordinary run of substances; for it a 10,000-fold decrease of speed is accompanied by a 25 per cent decrease of force instead

of a decrease by a factor of 10, as for tin. With mica, on the other hand, a 10,000-fold variation of speed is accompanied by no measurable variation of the force. Mica is more nearly typical of the great majority of substances.

It seems not unreasonable to suppose that some recrystallization mechanism may play an important part here. Doubtless recrystallization takes place, and it would be expected to be accompanied by a release of stress or by a flow at constant stress. Substances near their

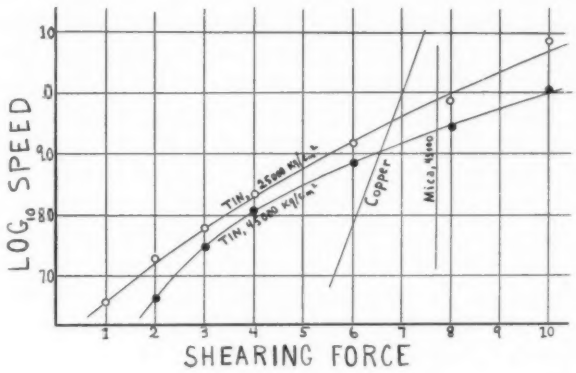

FIG. 3.—Shows speed of yield as a function of mean torque or shearing force. Both co-ordinates are in arbitrary units.

melting-points may recrystallize rapidly; substances farther from it, more slowly.

Substances vary in the way in which they flow under constant impressed force. The metals quite often harden. Under a constant force flow may start at a fairly high rate, gradually slow down, and eventually cease altogether. Silver is a good example. Mica, on the other hand, is an example of a substance that becomes weaker as a result of flow; a constant force greater than a critical value at first produces a low rate of yield, which increases until a point of instability is reached, where the weight runs away until stopped by the limits of the apparatus.

The phenomena of snapping or rupture already discussed may be importantly affected by rate of deformation. The frequency of snap-

ping seems to always increase with increase of speed of deformation. This is not merely an increased frequency in time, but the ruptures are separated by smaller angular displacements. Thus, in the case of borax glass an increase of speed of 60 fold carried with it a 3-fold increase in the total number of ruptures for a given angular displacement. Such an effect can only mean some mechanism of internal viscous readjustment operating during the displacement in such a direction as to prevent rupture.

One may speculate as to the possibility of rupture eventually ceasing altogether at low enough rates of yield. Extrapolations mean little, but borax glass starts off in a direction admitting the possibility that rupture will survive at infinitely low rates of yield. The fact that some substances rupture with no measurable plastic flow points in the same direction. On the other hand, substances have been found for which the phenomena of rupture entirely disappear at low-enough rates of deformation. The phenomenon is specific with the material.

The existence of these various time effects suggests crucial questionings with regard to behavior in geological times. We are here, of course, in a field of almost pure speculation; but there are examples of such abrupt freezing of internal mobilities with decreasing temperature or increasing pressure that I think we should not be frightened by the bogy of even geological time, but in our speculations freely use such concepts as shearing strength, admitting that the apparent strength in 10,000,000 years may be a little lower than for laboratory experiments, or the strength for 100,000,000 years a little lower than for 10,000,000, and also recognizing that the difference between geological and laboratory behavior may be a strong function of the material. We must, at any rate, recognize that the existence of these time effects means that no numerical datum or descriptive concept formulated in simple language is sharply applicable to geological phenomena.

There are other sorts of phenomena. There is almost always welding of the material. The initial material is usually a finely divided powder, compressed enough not to fall apart on handling. After shearing, this often emerges as a homogeneous disk, completely self-welded, and in many cases transparent if the original material was

naturally transparent. There is also in many cases welding to the steel. This is particularly prominent in the metals. Low-melting metals, like tin and lead, are as tightly welded to piston or anvil as if they had been soldered. Many non-metallic substances have no tendency to weld to the steel, but there are many striking exceptions. Borax glass is an example. A powder of borax glass is converted by shearing under pressure into a homogeneous transparent mass completely self-welded through its interior, and sticking to the steel itself as tightly as the familiar fused borax bead, actually wetting the steel around the edges. Calcite is another substance that sticks to the steel as if fused. These welding phenomena are exactly what would be expected. If the molecules are once brought into the range of each others' forces, they must attract just as if they had been originally neighbors in the same homogeneous piece of matter. Such intimacy of contact is favored not only by the magnitude of the forces but by the severity of the deformation under which any adsorbed surface films which might initially prevent contact must eventually get rubbed through.

Welding under the action of shearing forces must be of almost universal geological occurrence.

Permanent, non-reversible changes of density may be produced in some substances after exposure to high pressure and shear. Such changes may be particularly expected in substances which are capable of existing in two forms of different density. Thus chalcocite is permanently changed throughout its entire mass into the ordinary cubic Cu_2S. This was established by X-ray examination, for which I am much indebted to Dr. R. B. Jacobs. Quartz glass becomes denser, presumably partially assuming the crystalline condition. Calcite becomes denser, presumably partially turning into the denser aragonite. The changes of density are small and were determined by a floatation method. It is probable that in the two last cases the transformation is confined to a thin surface layer. This would be expected if the product of the transformation has a lower shearing strength than the original material, for when a thin surface layer of transformed material has once been formed flow may continue to take place in it, thus preventing the shearing stress in the rest of the disk from building up to the value which originally initiated the

transformation. This would definitely be expected in the case of SiO_2 mentioned above, because the transformation is presumably from a glass to a crystalline form. It has also been my experience that hexagonal crystals are particularly likely to show snapping phenomena, so that a higher shearing strength for the hexagonal calcite than for the rhombic aragonite would be consistent with other experience.

In general, plastic flow provides a means by which the effects of internal viscosity may be partially overcome, facilitating the occurrence of any transition which is demanded by the thermodynamic relations. A number of examples have been found of transitions which have run at room temperature under shearing stress which did not run at the same temperature under hydrostatic pressure. At 150° C. these transitions ran under hydrostatic pressure. There is no reason to think, however, that plastic flow is capable of completely overcoming viscosity, so that probably there are many cases among substances with high melting-points of transitions thermodynamically possible which even shearing stress cannot induce to appear. The transformation of graphite to diamond is probably an example.

Finally, there are chemical effects. These are difficult to study because of the very small quantities of material involved; not only is the total quantity small, but the chemically altered regions are usually confined to the surface layers. A few cases of chemical reaction have been definitely established. Bi_2O_3 is reduced to metallic Bi. This was proved by X-ray analysis, for which again I am indebted to Dr. Jacobs. Indirect evidence from the nature of the shearing curves makes it highly probable that there are a number of other cases of decomposition of metallic salts with liberation of a surface film of the pure metal. This is almost certainly the case for some other bismuth salts, for several of the salts of mercury, and probably for lead and others. The reduction need not go as far as a complete decomposition. SnO_2 is very definitely reduced to SnO, established by X-ray analysis by Dr. Jacobs.

Chemical synthesis may also be accomplished. Thus a mixture of metallic copper and sulphur is converted to Cu_2S, the product being a mixture of chalcocite and the ordinary cubic modification. The

part played by chemical alteration under shear may obviously be very important, indeed, in many phenomena of mineral or ore formation.

In conclusion, it seems to me that the nature of some of these experimental results not only may affect actual geological speculations about conditions deep in the earth but may affect also the tactics of such speculations. Some of the results are unequivocal enough, such as the great increase of resistance to plastic flow at high pressure, or the prevalence of polymorphism, or the existence of mechanical instabilities, to permit confident application to speculative problems. But there are other phenomena which present possibilities of such richness that the immediate effect of their discovery is likely to be embarrassment. Any speculation which hinges, for example, on whether a substance hardens or softens under prolonged plastic flow, or whether rupture in shear becomes more or less frequent at high pressures, must for the present be held in abeyance, for there are many examples of each of these sorts of behavior. Geology is rapidly approaching a point where an ultimate problem which has been staring it in the face can no longer be sidestepped, namely, to determine the actual specific physical and chemical behavior of those materials which do actually constitute the earth's crust.

ELONGATION IN DEFORMED ROCKS[1]

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ABSTRACT

Field data from a deformed clastic sedimentary series in northwestern Ontario disclose two directions of elongation—one indicated by the folding; the second, by the parallelism of linear elements, such as conglomerate boulders and individual mineral grains, to the fold axes. These two directions of elongation are normal to each other and are unequally developed, the major one being parallel to the linear elements and to the axial lines of the folds.

A hypothesis recently outlined by Sander best explains this dual elongation. It postulates a shear couple essentially normal to the axial lines of the folds, combined with intergranular rotational movements. The intermediate axis of strain (*B*) is parallel to the fold axes and to the linear elements and is unique in that it is a minor axis of elongation (for triaxial deformation) and also an axis of rotation. This results in the effective elongation of the structure being parallel to *B*, while elongation parallel to *A* (the long axis of strain) is retarded in its action.

INTRODUCTION

The problem of elongation in deformed rocks has received fresh impetus since the European methods of investigating rock fabrics have come into use.[2] The present paper introduces a field problem the interpretation of which hinges on these recent studies. It deals with the relations of folds to the planar and linear elements found in deformed rocks and is a direct field application of the data of structural petrology. The relations of planar elements (schistosity, cleavage, etc.) and folds are well known, although there has been little unanimity of interpretation. The linear elements (elongated grains, stretched conglomerate boulders, etc.), on the other hand, have not been studied in nearly the same detail; and their field relations to folds and planar elements have therefore not always received the attention they deserve.[3]

[1] Published with the permission of the Provincial Geologist, Ontario Department of Mines, Toronto.

[2] An introduction to these methods, the data obtained, and their present interpretation may be had in the following references: Bruno Sander, *Gefügekunde der Gesteine* (Vienna: Verlag Julius Springer, 1930); E. B. Knopf, "Petrotectonics," *Amer. Jour. Sci.*, Vol. XXV (1933); H. W. Fairbairn, *Introduction to Petrofabric Analysis* (Kingston, Ont.: Queen's University, 1935).

[3] The terms "linear" and "planar" element are useful both for their descriptive character and for their brevity. A foliation surface is composed of planar elements; a "cylindrical" or "pencil" structure is composed of linear elements. If associated, as is usual, the linear elements lie in the foliation surfaces. "Element" includes either individual mineral grains or a composite such as a conglomerate boulder.

Criteria of elongation in deformed rocks are given in Leith's *Structural Geology* (pp. 124–25) as follows: (1) distortion of pebbles of conglomerate; (2) distortion of mineral crystals; (3) distortion of volcanic textures; (4) distortion of fossils; (5) distortion of beds and attitude of folds; (6) relations to intrusives. The long axis of these distorted elements is interpreted as being parallel to A, the long axis of the strain ellipsoid. This picture of the structure is not applicable, however, to all cases. The writer recently had opportunity to study the relations of criteria (1), (2), and (5) over a considerable area in northwestern Ontario and from the evidence obtained concludes that the structural relations must be otherwise interpreted.

FIELD DATA

The Shabu sedimentary series in the vicinity of Birch Lake, Patricia district, Ontario, furnishes the data for the present discussion.[4] It consists largely of clastic material ranging from boulder conglomerate to fine graywacke and slate. Although the rocks have undergone intense folding and crumpling, primary structures are still preserved in most places, and the relations of bedding to the planar and linear elements of the formation are easily seen. The conglomerate boulders are distorted very conspicuously and form the most pronounced linear element. The accompanying sketch (Fig. 1) shows the structural relations in diagrammatic form. First to be noted is the relation of bedding and foliation, in which the foliation is parallel to the axial planes of the folds. This axial-plane foliation persists throughout the larger part of the sedimentary area examined. The second relation to be noted is the pitch of the folds, drawn here as vertical. It may vary as much as 20° from the vertical in different localities but is uniformly steep in character. The pitch is easily seen in the field (1) by the attitude of small drag folds and (2) by the trace of the bedding on the axial-plane foliation surfaces. The third feature of importance shown by the figure is the relation of the linear elements to the axial-plane foliation and the steep pitch. The linear elements consist of the conglomerate boulders and the individual mineral grains. The latter are arranged as linear elements on the foliation surfaces, and from the figure are seen to be parallel to

[4] W. D. Harding, "Birch Lake Area, Patricia District," *Ont. Dept. Mines*, Vol. XLV (1936).

the bedding trace. The former (conglomerate boulders) lie with their two longer axes parallel to the foliation surfaces in such a way that

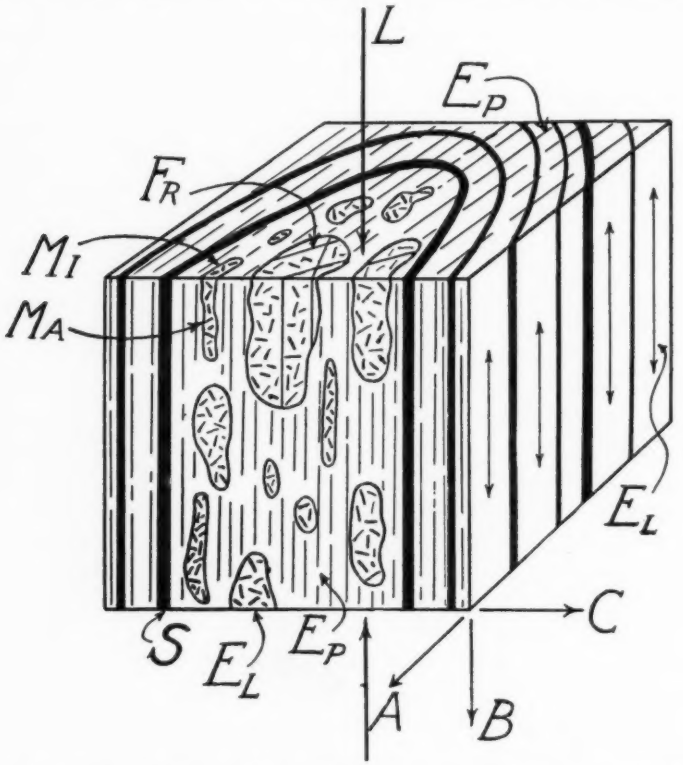

FIG. 1.—Schematic structural relations of Shabu sedimentary series, Patricia district, Ontario. *L*, axial line of folds (plunge, pitch, etc.); E_P, planar elements (schistosity, cleavage, etc.); E_L, linear elements (elongated grains, stretched conglomerate boulders; *S*, stratification; M_A, major dimensional axes of conglomerate boulders; M_I, minor dimensional axes of conglomerate boulders; F_R, shear fractures in conglomerate boulders; *A*, long axis of strain; *B*, intermediate axis of strain; and *C*, short axis of strain.

the longest axis of each (indicating the linear element) is parallel to the pitch of the folds. We have, therefore, two types of linear elements which are parallel both to the foliation and to the known pitch

of the structure. A fourth feature is the relation of fractures to the linear elements. Steeply dipping shear fractures are found in many of the conglomerate boulders; if two sets are present, their line of intersection is parallel to the pitch of the linear elements.

There can be no question that the direction parallel to which these elements are arranged is the major direction of elongation. The field evidence supplied by the conglomerate boulders is sufficient in itself. In many instances massive granitic boulders have one dimension ten to fifteen times greater than the least dimension. Fragments of previously foliated rocks might have unequal dimensions of this magnitude before their second deformation as part of a conglomerate formation, but certainly not material of originally massive texture such as granite.

ORIGIN OF STEEP PITCH

The origin of the steep to vertical pitch of the structure just described may be considered independently of the elongation. Two possibilities are: (1) tilting of the beds contemporaneously with deformation and (2) cross-folding. The first assumes that in the very first stage of deformation of the unfolded sediments the axial lines are horizontal. As folding and deformation progress, however, tilting may occur as an external movement unrelated to the internal deformation of the sediments. This process would result in a steep pitch such as described.

The second possibility, cross-folding, assumes a regional deformation in which isoclinal folds are formed whose axes are essentially horizontal. A second, more local deformation follows, causing movement to occur parallel to these fold axes. These secondary fold axes lie therefore at steep angles with the horizontal. Cross-folded structures of this type are well known. In the Shabu series and adjacent formations there is at present no direct evidence of either of these two modes of origin of steep pitch, and the elongation problem is therefore discussed independently of it.

TYPE OF DEFORMATION

Although the exact nature of rock deformation is at best only conjectural, two contrasting cases may always be considered: "plastic" deformation and "plane" deformation. If only the linear elements in the Shabu series are considered, it is conceivable that "plastic"

deformation could have brought about their present arrangement. For example, Mead[5] produced folds by deformation of wax covered with tinfoil—an essentially plastic deformation, since no shear ruptures occurred under the conditions of the experiment. The major elongation was definitely parallel to the axial lines (linear elements) since the tinfoil ruptured in several places normal to the fold axes. A minor elongation parallel to the axial planes of the folds and normal to the axial lines is shown by the arching up of the wax into folds. The strain ellipsoid of rupture has no application in deformation of this type since failure does not occur parallel to plane surfaces.

Two factors argue against a purely "plastic" deformation of the Shabu series: (1) the rocks possess pronounced planar elements in addition to the linear elements; (2) the arrangement of the conglomerate boulders parallel to the axial lines does not mean that their unequal dimensions are necessarily due to "plastic" deformation. Regarding (1), it might be argued that a "plane" deformation, subsequent to a "plastic" deformation, caused development of the planar elements quite independently of the elongation. The orientation of both major and minor directions of elongation of the conglomerate boulders *parallel* to the foliation planes, however, can scarcely be regarded as fortuitous, and points to a common origin for both linear and planar elements. Furthermore, the foliation is parallel to the axial planes of the folds, a circumstance which also indicates a common origin for both.

Regarding (2), a "plastic" deformation is permissible which affects only the less competent matrix and causes a dimensional orientation of the longer dimensions of the more competent conglomerate boulders parallel to the fold axes. As already indicated, these longer dimensions undoubtedly represent actual elongation and not merely initial shapes of the unconsolidated material, so that even if a selective "plastic" deformation is assumed, the elongation problem is still unsolved. From consideration of the available data, therefore, "plane" deformation is probably much the more important in this case, and further discussion is based on it alone.

[5] W. J. Mead, "Notes on the Mechanics of Geologic Structures," *Jour. Geol.*, Vol. XXVIII (1920), pp. 505–23.

NUMBER OF DEFORMATIONS

Two cases may be considered with respect to the elongation: either a single continuous deformation or two separate deformations. Assuming one deformation, the parallelism of the linear elements (indicating the direction of elongation) to the axial line, or pitch, of the folds appears mechanically impossible at first glance. In Figure 1 the elongation of the conglomerate boulders is parallel to a vertical axis; the elongation indicated by the folded beds is parallel to the horizontal axis lying in the axial-plane foliation. Is it evidence of two deformations, or may such structures result from one deformation?

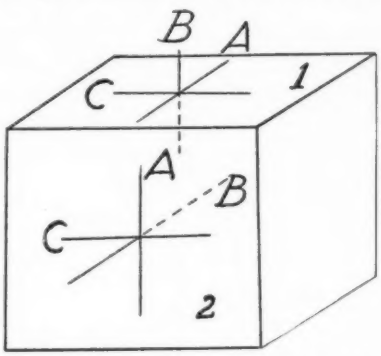

FIG. 2.—Showing the relation of the strain axes if two separate deformations are assumed. Compare with Figure 1.

The case for two deformations may be considered as follows: One deformation will be shown by the minor elongation of the conglomerate boulders (horizontal planes of Fig. 1); the other by the major elongation denoted by the linear elements (vertical plane of Fig. 1). Elongation is assumed to be parallel to *A*, the long axis of the strain ellipsoid. In Figure 1 the orientation of the strain axes as given would produce the minor elongation only (deformation No. 1); to produce the major elongation a deformation ellipsoid with *B* and *A* reversed would be necessary (deformation No. 2) (Fig. 2). *A* would then be vertical, *B* horizontal, and *C* fixed, with reference to the figure. For both deformations the same axial-plane foliation would develop, since the direction of shortening *C* is constant. The differences to be expected would be in the disposition of the fracture surfaces and drag folds, both commonly occurring evidences of deformation in these rocks. For deformation No. 1 the common intersection line of all fracture surfaces and the axial lines of all drag folds would be vertical (parallel to *B*); for deformation No. 2 these lines would be horizontal (parallel to the new position of *B*). There is field evidence

of *vertical* intersection lines and axial lines only; and deformation No. 2 would therefore appear to have been an earlier one, evidence of whose fracture surfaces and drag folds has been destroyed by the later deformation, No. 1. Deformation No. 2, however, produced the vertical elongation of Figure 1, which is the major elongation of the whole structure. From the evidence of the elongated boulders it was a much stronger deformation than No. 1; yet no trace is found of fracture surfaces and drag folds which can be ascribed to it. The case for two deformations conflicts, therefore, with the available field evidence. Furthermore, the presence of axial-plane foliation indicates the probability that the secondary structures are dependent on the configuration of the primary structures. That is, it is probable that the folding and development of cleavage were essentially contemporaneous in order to produce the symmetrical axial-plane relations observed in these structures. Further discussion is based, therefore, on one continuous deformation.

HYPOTHESES OF ELONGATION

Two hypotheses based on one deformation may be used to explain elongation parallel to the axial lines of folds. The first is deductive and is based on Mead's experiments[6] in which the long axis A of strain coincides with the axial line. The second, formulated by Sander, is inductive and places the intermediate axis B parallel to the axial lines. Mead's well-known experiment with brittle paraffin is represented in Figure 3. A shear couple (shown by the arrows) produces elongation parallel to the long diagonal of the parallelogram and shortening normal thereto. Shear fractures develop parallel to the sides of the parallelogram. This type of failure requires B to be normal to the plane of Figure 3 and A parallel to the long diagonal. The principal feature of this hypothesis is that the shear couple is parallel to pre-existing, potential planes of slip in the structure, and that the axial lines of the resultant folds (parallel to the long diagonal) lie in the plane of the shear couple.[7] Although the

[6] *Ibid.*

[7] This is equivalent to a so-called "horizontal" shear couple. It is better expressed with reference to potential planes of slip only, since it may apply to structures previously deformed in which potential planes of slip need not be horizontal. Similarly, Sander's hypothesis postulates in part a so-called "vertical" shear couple.

hypothesis is mechanically quite feasible, it cannot be applied on a regional scale to any deformed area as yet studied. Chamberlin[8] has already pointed this out in the case of Appalachian structures. In these the major elongation is in general parallel to the axial lines of the folds, but the fault planes are also in general parallel to these axial lines. Mead's experiment involves shearing which acts obliquely to the axial lines. Faults having this orientation are lacking in the Appalachians, except perhaps locally. This is also true of Alpine structures. Sander's hypothesis,[9] now to be described, is an attempt

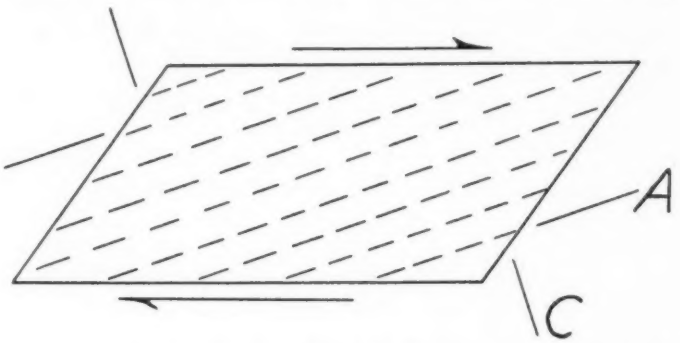

FIG. 3.—Diagram of strain axes in a body subjected to a shear couple acting parallel to the potential slip planes (so-called "horizontal" shear couple).

to explain the origin of this type of structure in which the axial line of the folds is parallel to shear surfaces.

In contrast to the deductions made from Mead's experiments, Sander's hypothesis is twofold and postulates (1) a shear couple normal to the potential planes of slip, and therefore normal to the axial lines of the folds, and (2) intergranular rotational movement of the particles affected by the stress. The effect of this combined movement in the example described from the Shabu series (Fig. 1) would be as follows: The fold axis represents *B* (intermediate axis) of the strain ellipsoid; the direction in the axial-plane foliation normal to *B* represents *A* (the long axis). *C* is normal to the axial-plane foli-

[8] R. T. Chamberlin, "The Strain Ellipsoid and Appalachian Structures," *Jour. Geol.*, Vol. XXXVI (1925), p. 85.

[9] *Op. cit.*

ation. This orientation of the strain axes obviously explains the elongation with respect to the folds, in which elongation and shortening are represented by A and C, respectively. The apparent anomaly with respect to the elongation parallel to B, shown by the linear elements in Figure 1 (conglomerate boulders and grains), has its explanation in external rotational movement of the particles about B as an axis. The main direction of movement of material during the deformation is in, and parallel to, the AC plane of the strain ellipsoid. This is true both for shearing movements (related geometrically to the strain ellipsoid) and for the intergranular rotational movements (independent of the strain ellipsoid). The strain axis B is unique, however, in that the major elongation of the elements is parallel to it. This is due in part to the geometrical properties of the strain ellipsoid and in part to the combined shear and rolling movements postulated above. We assume first a triaxial strain,[10] or deformation, i.e., one in which all three axes of strain change in length during the deformation, so that $(A+C)/2 < B$. (In a biaxial, or plane, deformation B remains of constant length, so that $[A+C]/2 = B$). A and B are both elongated, A more than B, in order that the relation $A > B > C$ may hold, the associated shearing movements taking place mostly between the beds and parallel to the AC plane of strain. The independent rotational movement about B as an axis modifies the shearing movements as follows: The mineral grains are rotated about B as a fixed axis, the direction of movement being parallel to the AC plane of strain. This intergranular rotation does not permit continuous elongation of the grains parallel to A to take place, i.e., a given direction in a grain lies parallel to A at one period and is exposed to elongation; rotation, however, is more or less continuous, and this same direction in the grain is later subjected to shortening parallel to C. Thus the elongation effect parallel to A is not fully expressed. With respect to B, the same direction in a grain is continuously parallel to this axis during the whole deformation (since it is the rotation axis); moreover, in triaxial deformation it is an axis of elongation, so that the grains undergo continuous elongation parallel

[10] Triaxial strain is assumed since it is a more general case than that represented by biaxial. The latter involves an invariable confining stress (so that B remains constant), which in nature is probably seldom realized.

to it. The combined effect of the shearing and independent intergranular (rotational) movements is such that, although *B* is only a minor axis of elongation, it is the most effective one in the rock structure.[11] This is shown in Figure 1 by the parallelism of the linear elements to *B*. The conglomerate boulders have their major axes of elongation parallel to the fold axes; their minor axes lie in the foliation parallel to *A*. Without the assumed rotation around *B*, the principal elongation would be parallel to *A* rather than *B*.[12]

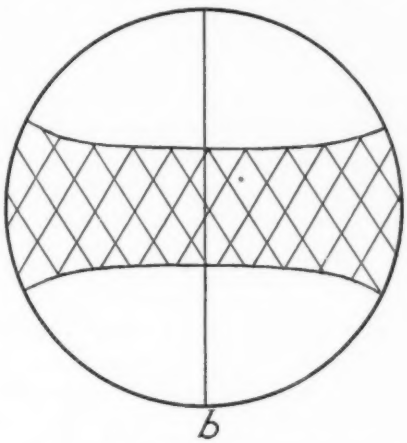

FIG. 4.—Orientation zone (cross-hatched) of vertical crystallographic axes of quartz. The zone is normal to *b*, the direction assumed by the parallel linear elements.

Although no microscopic study of mineral orientation was undertaken for the present problem, this conception of intergranular rotation combined with a shear couple normal to the axial lines is based on such data. The typical orientation of the vertical crystallographic axes of quartz plotted on an equal-area net is such that a girdle of axes lies normal to the parallel linear elements. The relation of such a girdle to Figure 1 would be as shown in Figure 4, in

[11] This independent rotation is Sander's external rotation (*Externrotation*) and is not to be confused with "internal rotation of the strain axes" (*Internrotation*). The latter was first described by Becker ("Finite Homogeneous Strain," *Geol. Soc. Amer. Bull. 4* [1893]) in his analysis of rotational strain, in which he showed that a slight angular rotation of the strain axes with respect to the chief stress would result in unequal development of the shear surfaces. Quantitatively this internal rotation is inconsiderable, however, and would not sensibly affect the present hypothesis. The strain axes may therefore be regarded as fixed.

[12] This has been found to be the case in certain shear zones in which the elongation is normal to *B*. This may arise from a local, relatively high velocity of deformation in which the shearing movements are quantitatively much greater than the intergranular rotational movements. Thus the effective elongation would be parallel to *A* instead of *B*.

which *b* is parallel to the linear elements and to the axial lines of the folds. Within this girdle, moreover, are concentrations of axes which have a definite relation to the foliation surfaces of the rock. With regard to the hypothesis of elongation just outlined, it is believed that the local concentrations of axes represent largely the effect of strain and that the girdle as a whole is the locus of axes which show the effect of intergranular rotational movement about the axial lines. This picture fits with that of combined shearing and rolling movements about the intermediate axis *B* of the strain ellipsoid. Quartz is ordinarily the chief mineral the orientation of which is determined (owing to its abundance and sensitivity to deformation), and the universality of its general type of orientation (Fig. 4) thus makes an interpretation possible on the basis of field data alone.

CONCLUSION

Although attention has been focused in the preceding discussion mainly on a small area, the problem presented is not a local one. Every deformed structure has directions of elongation and shortening, and field study of its linear and planar elements is fundamental. The hypothesis just presented explains, in the writer's opinion, more of the facts than any other. It is to be hoped that succeeding studies will bring to light new data and fresh points of view.

ACKNOWLEDGMENTS.—The field study arose in the course of geological exploration undertaken by the Ontario Department of Mines; and the writer is indebted to W. D. Harding, in charge of mapping, for frequent discussions of the problem. The manuscript has been critically read by C. H. Behre, Jr., Marland Billings, E. L. Bruce, David Griggs, W. D. Harding, J. E. Hawley, Andrew Leith, and C. O. Swanson, to all of whom the writer acknowledges his appreciation.

QUATERNARY GLACIATION IN THE COAST RANGE, NORTHERN BRITISH COLUMBIA AND ALASKA

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ABSTRACT

On the higher mountains in the Coast Range of British Columbia and Alaska the action of ice and snow is so severe and the surface is changing so rapidly that little record of old surface features remains and an estimate of the amount of erosion by ice is difficult to make. In the valleys likewise, ice erosion has been active and the surface has changed rapidly; but here reference to certain datum planes, such as sea-level and the levels of certain rivers, suggests that glacial scour has deepened some of the major valleys 2,000 feet or more.

In the period between 9,000 and 4,500 years ago conditions of intense alpine glaciation prevailed. Between 5,500 and 4,500 years ago the climate was moderating; the ice retreated rapidly and disappeared from Stikine and Taku fiords. From 4,500 until about 1,500 years ago the climate was warmer than at present. Subsequently the ice advanced in places for 20–25 miles beyond present glacier termini. It then retreated to its present stage.

INTRODUCTION

Most peaks above 5,000 feet in the Coast Range of British Columbia are surficially shattered, as if by dynamite. The fractured rock is continually falling away or being carried down by ice. Great talus slopes and moraines flank many peaks, and large quantities of débris derived from the peaks are distributed along the glaciers and at their termini. Sufficient erosion has taken place in the high mountains since they were uncovered by the retreat of the last ice sheet to give an entirely new land surface and therefore destroy much of the evidence of Pleistocene glaciation. In fact, any reliable estimate of the amount of ice erosion for the whole of the Quaternary or a deciphering of the Pleistocene history of the region is impossible. The effectiveness of the more intense ice stages therefore must be gauged by the erosion in the valleys where some data are available.

Many large valleys, though clearly fashioned by glaciation and in places still occupied by glaciers today, are deeply filled with débris and therefore are not subjected to ice erosion. While a few of the larger valleys in their lower reaches and many near their heads are occupied by ice which is cutting bedrock, glacial erosion in the valleys at the present time is, in general, small compared with that of

more extensive ice stages. The effectiveness of ice erosion in valleys therefore is not shown by present activities so much as by the results already attained.

Four ice stages are recognized in the Coast Range of northern British Columbia: (1) "alpine stages" like the present, in which ice movement largely followed the gradients of the valleys; (2) "intense alpine stages," when valley glaciers completely occupied trans-range

FIG. 1.—Looking south across Talsekwe Valley in the foreground and Taku Valley on the left. In the foreground the rock even on fairly flat surfaces is highly shattered. In the background many great talus slopes, in places thousands of feet long, testify to rapid removal to lower levels of the shattered blocks. Under such conditions it is clear that the surface everywhere is rapidly changing and that no evidence of old surfaces is likely to be retained. Photograph by International Boundary Survey.

valleys and in the eastern part of the range advanced up the slope into the interior; (3) "mountain ice-sheet stages," when the range was buried in ice and movement was in both directions from the axis of the range; and (4) "continental ice-sheet stages," when ice moved westerly across the whole range.[1] The effectiveness of glacial erosion in each stage varied greatly in different parts of the range. In the western half of the range erosion was active during all stages

[1] Forrest A. Kerr, "Glaciation in Northern British Columbia," *Trans. Roy. Soc. Can.*, Sec. IV, Vol. XXVIII (1934).

in practically all parts, except in the great valleys during alpine stages. In the eastern part ice was not sufficiently extensive during alpine stages to cause much erosion, and during continental stages the westward ice movement in the valleys and along lower parts of the mountains was probably greatly reduced and erosion restricted because the range acted as a dam. Most of the erosion in the eastern part therefore was during the intense alpine and mountain ice-sheet

FIG. 2.—The Coast Range near Stikine River. Here again it is apparent that the surface is changing so rapidly that no accurate estimate of what it was even a few thousand years ago can be made. Under such conditions no accurate estimate of the extent of erosion over any long period can be made. Photograph by International Boundary Survey.

stages, when the ice moved eastward. The minimum ice erosion in the great trans-range valleys then was near the axis of the range and just beyond the limit of the mountains, both to the east and west, where a spreading-out of the ice reduced its downward pressure and diminished its erosive power.

At the present time some fiords, such as Tracy Arm and Glacier Bay, are occupied by great glaciers discharging into the sea. These have but recently retreated from full occupancy of the valleys. In these fiords the water is very deep, the valley walls are precipitous,

and there is a marked scarcity of unconsolidated materials. Above sea-level, similarly, valleys are clearly indicated to have been heavily eroded by ice. It is believed therefore that all valleys when occupied by active glaciers of great thickness are kept scoured to bedrock. Thus, during the last period of intense alpine glaciation little, if any, of the earlier unconsolidated deposits would remain in any major valley, even in the trans-range valleys such as the Taku and Stikine. The filling now in the valleys therefore records only the period since the last intense alpine stage. Just within the eastern limits of the range and beyond in the interior plateau, however, conditions were quite different, and extensive unconsolidated deposits formed during the Pleistocene still remain. No study has yet been made of these.

EXTENT AND DURATION OF ICE EROSION INDICATED IN TRANS-RANGE VALLEYS

The withdrawal of the great glaciers from Stikine and Taku valleys marked the end of the period of erosion here and of the last intense alpine stage. It also marked the beginning of deposition in these valleys and therefore the beginning of the record of recent geological history.

Data from Stikine Valley.[2]—In the axial part of the Coast Range, between the tributaries Butterfly Creek and Deeker Creek, the rock walls of the Stikine Valley are so close together at several points that the gravel fill cannot be very great. Just below Deeker Creek rock is exposed most of the way across the valley, indicating that the glaciated base, where covered, cannot be far below the present gravel surface. Since the present river channel here has an elevation of about 250 feet, the glaciated base can be safely considered as 200 feet above sea-level. Deeker, Ochsa, and Butterfly creeks enter Stikine Valley through deep canyons cut by postglacial or interglacial streams in the lips of hanging glaciated valleys. Allowing for no other Quaternary deepening of these valleys than the cutting of canyons, the projection of the glaciated bases and the study of the general topography clearly indicate that the old Stikine channel could easily have stood at 300 feet above sea-level, so that the estimate of 200 feet is probably a conservative minimum.

[2] "Stikine River Area," *Can. Geol. Surv. Maps 309A, 310A, 311A* (1935).

Above Butterfly Creek, Stikine Valley is wider than below; and it continues to widen out, especially above the junctions of large tributaries, as far north as the outer limit of the range, beyond which it narrows. Here the elevation of the base is 500 feet. Above this,

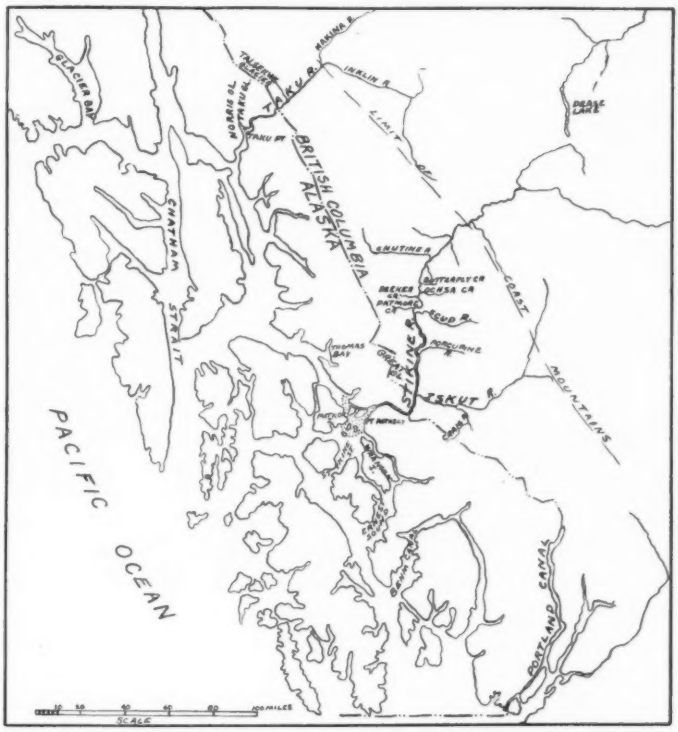

FIG. 3.—Map of the Coast Range Area of British Columbia and Alaska

along the walls of the valley, are remnants of Pleistocene and possibly older lavas which originally filled the old valley. In places they extend practically to the bottom, indicating that the old filled valley had a base at approximately the same level as the present one. Thus the preglacial valley may be considered to have stood at 500 feet above present sea-level at the eastern limit of the range and at

300 feet at Butterfly Creek. The projection of this gradient to the outer limit of the Alaskan islands (assuming the straits and bays are due to overdeepening by ice) gives a reasonable gradient of about 2 feet to the mile for a river which must have been well established in its trans-range course long before the Quaternary. The projection of the gradient to the present mouth of the river gives equally reasonable results. Thus, in either case the pre-Pleistocene rock valley

FIG. 4.—Lower Stikine Valley. Photograph by Alaskan Aërial Survey Expedition of the U.S. Navy Department.

was not greatly different in elevation from that of the present gravel surface.

From Deeker Creek south, Stikine Valley widens; and below the junction of the Scud, a very large tributary, the rock walls are 2 miles apart and the valley base is entirely filled with gravel to an elevation of about 200 feet above sea-level. The topography indicates that the volume of ice from Scud and Patmore valleys was greater than that from the upper Stikine during those periods of glaciation when ice in the eastern part of the range moved eastward. Overdeepening of the Stikine at and below the Scud was therefore very great.

A comparison with other similar junctions which are not gravel-filled indicates that the rock base of the Stikine Valley probably lies much more than 200 feet below the gravel surface and is therefore below sea-level. Below the Scud the base must be cut as low or lower. Thus, about 80 miles of Stikine Valley are considered to be cut below sea-level. This is less than the length of Portland Canal and is clearly a reasonable estimate.

Below the Scud, Stikine River flows over gravel, sand, and silt with a fairly even, though gradually decreasing, gradient to the outer limits of the present delta. The valley filling consists largely of morainal material, outwash from glaciers, and delta, fluvial, and lake deposits. Advance of glaciers over deltas, as well as other changes, has made the filling process complex.

Any time estimate on the basis of the rate of filling must be somewhat inaccurate because of differences in rate due to climatic changes. During much of the filling period the climate was probably much warmer than at present; ice was therefore more limited, erosion less, and the supply of débris decreased. At other times during colder spells the supply was probably greater than at present. The building of individual tributary deltas and the effect of moderate advances of ice probably had little influence upon the general rate. On the other hand, the building-up of the gradient with a maximum increase in height of 200 feet at the Scud almost certainly has tended continually to decrease the load of the main river and tributaries, and thereby to cause a general decrease in rate of deposition over the total period of filling. Another factor is that of deposition during the period of retreat of the ice in Stikine fiord. Observations on modern glaciers show that retreat in similar fiords may be as fast as 12 miles in 40 years. If the retreat in the Stikine Valley was similarly rapid, the amount of deposition during this period would not be important; but the retreat may have been much slower.

There is little accurate information about the rate of modern delta growth. In 1793 Captain Vancouver's men mapped the coast. The vicinity of the Stikine River is shown with a fair degree of accuracy, and the delta is well defined. Soundings are shown at several points across the front of the delta, and apparently this map fairly well locates the delta front at that time. (The outer line marked on

the map is considered to be just inside the 2-fathom depth.) In 1886–87 the United States Coast and Geodetic Survey mapped the delta, providing the only accurate map available. During the 94 years between the two mappings the delta must have advanced at least $1\frac{1}{3}$ miles over a front of about 7 miles from Mitkof Island south to the mainland, which includes the greater part of the delta front. Probably the advance was nearer 2 miles, and it may have been even more. The present position, when compared with that of 1886–87, also indicates a rapid rate of advance. The estimated rate therefore is $\frac{3}{4}$–1 mile in 50 years, or 75–105 feet per year.

The delta at present is building among the coastal islands beyond the end of Stikine Valley proper. It is advancing over a front of more than 8 miles in water which has a maximum depth near by of about 600 feet. The Stikine Valley for the lower 40 miles to the Iskut River averages $2\frac{1}{2}$ miles in width, and the depth to its rock base is roughly estimated to be greater than that at the delta front in inverse ratio to the widths of the channels at these points. This would make a rough correspondence in the total extent of cross section of the unconsolidated materials within the valley and at the delta front. It is assumed that the rate of supply of material as represented by the present rate of delta accumulation is probably fairly representative of that during filling within the valley. It should be noted that the rate of advance now on the south side toward Wrangell Island appears to be greatly retarded by sea currents between the island and the delta front. Furthermore, such sea currents, since the delta advanced beyond the valley proper at Rothesay Point, may have been responsible for sufficient transport of material farther away to greatly retard present delta advance compared with that of the filling in the valley.

At a rate of 100 feet per year, the filling of the section from Iskut River to the position shown on the Alaska maps of 1886–87 would have taken over 2,100 years. Assuming a like rate[3] for the 40 miles from the Scud to the Iskut, the total time of filling becomes approximately 4,300 years. This figure, or roundly 4,500 years, is believed

[3] A like rate is used on the assumption that the decrease in supply of material above each tributary would be offset by a relative decrease in size of valley.

to represent a mean and is used as a basis of computation, though it must be recognized that it may be considerably in error.

Data from Taku Valley.—Vancouver's map of 1791 shows an embayment where Norris Glacier enters the inlet of Taku Valley. If the outline of the coast as shown is approximately correct, the advance of the delta and outwash plain of this glacier has been 2–3 miles. This is within the limits of possibility, but no definite value can be attached to these figures. Vancouver's map does not even suggest the approximate front of Taku River delta, so that it is of little help in this respect. The first accurate map of the delta was that by the United States Coast and Geodetic Survey in 1890. No later accurate data are available; but in the fall of 1932, observations of stranded icebergs and data supplied by William Strong, who had navigated the river for the previous 15 years, were used for sketching the approximate outline of the delta. The map of 1890 shows the main delta front trending approximately north from a point about 1 mile northeast of the tip of Taku Point. Upriver from this for an average distance of a mile there was a large area with depths of 5–12 feet below mean low water. In 1932 it appeared that this whole area, except for channels, had been filled in, so that the depth was in the main much less than 5 feet. Furthermore, the whole front of the delta appeared to have moved forward, so that it extended about due north from the northeast protuberance of the main part of Taku Point. It is believed that the advance of the main front of the delta (represented on the map by an abrupt drop of 5 fathoms or more) has been at least 2,000 feet over a front of 10,000 feet or more. Mr. Strong had noted over the 15 years prior to 1932, and particularly during the last 6 years of this period, a very rapid advance of the shoals. It is to be remarked, of course, that the inlet here has depths ranging only up to 230 feet, and it seems probable that it has already been partly filled. It may also be somewhat shallower, as far as the rock base is concerned, because of the junction here of the three glaciers during more extensive ice stages. Further, the vast inpouring of débris from Norris Glacier, just to the west, might be an important factor in building up the floor, while the tide waters running in and out of the bay in front of Taku Glacier undoubtedly carry an unusually large amount of material farther away.

It is impossible to evaluate properly all these influences. There is no doubt, however, that the rate of advance of the delta is rapid and now approximates 1 mile in 100 years. It seems altogether likely that this rate will be maintained and possibly exceeded during the next 10 miles, since filling is already far advanced for four of these by Norris Glacier. A large lake will be left in front of Taku Glacier,

FIG. 5.—Taku River. On the left the filling of the valley above the base of the glacier and subsequent retreat of the ice caused the formation of a basin now occupied by a lake. Photograph by the Alaskan Aërial Survey Expedition of the U.S. Navy Department.

and upon the extent to which this is filled will depend the rate of advance of the main delta.

It is estimated that the rock base of all of Taku Valley is cut below sea-level, so that the fiord at one time probably occupied practically all of the valley. Thus the advance of the delta for the 45 miles of valley would have required 4,500 years. It is a remarkable coincidence, though this cannot be taken as a criterion of correctness, that this figure is of the same order as that deduced for the filling of Stikine fiord.

Data from Portland Canal; Conclusion.—Hanson,[4] from his work in Portland Canal, estimates that the Bear River delta has been built in 2,500 years. The age of all three deltas may be approximately the same, though this is not necessarily so. The extent to which the fiord sections of the three valleys are filled shows a considerable range. The extent of present ice filling in fiords where glaciers reach the sea is also very different. There are many factors in the history of the valleys which could explain these differences other than the time of retreat of the glaciers. Since Taku and Stikine valleys show similarities in present glacier positions, filling, and in other respects, it is thought that 4,500 years in each is a fair tentative working-figure for the period since filling began and since the end of the last intense alpine stage.

HISTORY IMMEDIATELY PRIOR TO 4,500 YEARS AGO

Since accurate data of the Quaternary history prior to this last period are not obtainable in the Coast Mountains, it is instructive to make some comparison with other sections. Antevs[5] states that by 9,000 years ago the ice sheet in Eastern North America had retreated to the Labrador Peninsula. If changes in climatic conditions were general throughout North America, it seems reasonable to assume that, if an ice sheet existed in the Labrador Peninsula, at least intense alpine conditions[6] prevailed in the Coast Mountains. Similarly, during more extensive ice-sheet stages in the east, it would be natural to expect ice-sheet conditions over the Coast Range. Thus it is assumed that between 9,000 and 4,500 years ago intense alpine conditions prevailed.

COMMENCEMENT OF THE WARM PERIOD

Antevs states that in Sweden the summer temperatures were higher than at present during a period from about 7,000 to 3,000 years ago. A similar warm-climate (or possibly drier-climate) period occurred in the northern Coast Range section. The Great Glacier

[4] George Hanson, "Salmon River Area," *Can. Geol. Surv. Mem. 132* (1922).

[5] Ernst Antevs, "Late-Glacial Correlations and Ice Recession in Manitoba," *Can. Geol. Surv. Mem. 168* (1928).

[6] During these times ice fields must have been much more extensive than now, and probably in places approached ice-sheet conditions.

FIG. 6.—The Great Glacier debouching into Stikine Valley. Note the two distinct moraines on the right. Shells and logs occur in the moraine at the ice front in the section between the large stream in the left center foreground and the center of the crescent lake on the right.

which extends out into the center of Stikine Valley carries sea shells and large pieces of wood which must have come from mature trees. There is no evident source for such materials in the present drainage basin. Undoubtedly they have been picked up recently by the ice in its course. Therefore, at some recent time the sea, standing higher than at present, extended up the Stikine Valley and probably up Great Valley. Shells were deposited on beaches, and probably mature trees grew in Great Valley where now there is only ice.

FIG. 7.—In Talsekwe Valley aggradation is so rapid that a wall is being built across the tributary valley on the right, and soon a lake may be formed there. Marine shells and mud were washed by floods from under the glacier and were later found after thawing in mounds on the outwash plain. Photograph by R. Bartlett, Geological Survey of Canada.

From beneath Talsekwe Glacier in the Taku district great volumes of water escape periodically, owing to the bursting of an ice-dammed lake. These outbursts cut deeply into the gravel bed below the glacier. In 1929 an outburst brought out frozen cakes of mud carrying sea shells. In 1932 the outburst laid bare alder stumps and moss, *in situ*, 15 feet below the surface of the frozen gravel which previously underlay the tip of the glacier. Thus the sea had extended beyond the end of the present glacier, which is 400 feet above sea-level, and shells were deposited on the beaches. Probably at some later date

the sea had retreated and alders grew where the ice is now. This evidence indicates a period of warm climate of considerable length.

Along this portion of the Pacific Coast the sea is known to have stood at least 600 feet higher than at present. Since the marine deposits of Talsekwe Valley at 400 feet or higher were laid down after the retreat of the ice 4,500 years ago, the sea must have stood 400–600 feet higher at that time than now. Marine deposits in the Stikine and Taku valleys are very scarce, indicating that they were probably of limited extent. Thus the time during which the sea stood higher in these valleys must have been relatively short. Antevs estimates that the land rose about 600 feet in 2,000 years after the retreat of the ice.

There is no evidence of depression of the land under the present ice burden or the much greater burden of the recent advance that put much ice in Stikine Valley. Therefore it seems likely that sufficient removal of ice load to permit rising of the land would have occurred even before the complete retreat of ice from the Stikine Valley, and therefore before the end of the last intense ice stage 4,500 years ago. (There may, of course, have been some lag in resilience.) The fact that the sea stood at 400 feet or higher after the ice had retreated even farther than its present limit, which is well back of its position at the end of the last intense alpine stage, indicates a very rapid retreat, compared with more recent retreats to be described later. Pronounced moderating of the climate would seem therefore to have been responsible for the rapid retreat and to have begun sometime before the water reached the 400-foot level and probably before the land started to rise. A recent relatively short retreat of ice is estimated to have taken about 1,000 years; so the period of the more rapid but more extensive retreat is tentatively placed between 500 and 1,500 years. On this basis the warm climate began 5,000–6,000 years ago. From the foregoing data and Antevs' estimate of the rate of rise of the land it is estimated that the sea started to recede between 5,000 and 4,500 years ago and that it reached its present position before 2,500 years ago.

END OF WARM PERIOD

The end of the period which was warmer (or possibly drier) than the present climate was marked by an advance of the ice to positions

corresponding to or beyond those at which it now stands. The alder stumps and moss *in situ* below Talsekwe Glacier suggest that the warm period lasted until after the sea had retreated. The rate and extent of aggradation, however, afford data for an estimate of the time; and this can be checked against an estimate of time required for the retreat of the glaciers from their most advanced stage (of the recent advance), which is based on other data.

Computation of the rate of aggradation is greatly complicated, first, by the influence of the early delta deposits which were built far above present sea-level; and second, by the advances of glaciers in recent times. Total aggradation of 200 feet occurs in Stikine and Taku valleys at the heads of the fiord sections. As these are the starting-points for delta-building and aggradation, the total thickness of 200 feet represents 4,500 years (or slightly less) of valley filling, thus giving a rate of 1 foot in 20–25 years. At the present time there are many stretches where aggradation undoubtedly is much faster than 1 foot in 20–25 years; but in others it is much less, and in some it is absent.

The Great and Taku glaciers still ride over deposits of shells, mud, and sand. It is therefore clear that they have not greatly modified their channels since advancing over these unconsolidated materials. Some of the shell-bearing material has been cut into, but probably not to any great extent. Also, it is clear that the valleys must have been well filled before advance of the ice, not only under the ice but beyond, for otherwise there would have been such a steep gradient that either streams or the glaciers would have cut into the unconsolidated materials much more than they have. Therefore, it is evident that filling was largely completed before the advance of the ice. Further, the presence of the shells at the top of this filled base below the glaciers indicates, as already stated, that the sea was at a high-level stage comparatively recently and during the later stages of the filling, for otherwise the shell beds would have been cut down to lower levels and later covered by aggraded materials. These data also support the postulate of a short period since the recession of the sea, and therefore since the retreat of the ice.

Though the glaciers have not aggraded their channels beneath the ice, débris from them is responsible for much aggradation beyond.

The river channels throughout their lengths show an even gradient, steepening slightly from the deltas to the ice fronts. This indicates that the aggradation has occurred throughout the full lengths of these channels and not just in front of the glaciers. Holes in front of the ice and channels cut in the outwash indicate that the base of the ice of Great, Talsekwe, and other glaciers is 50 feet or more below the top of the outwash plain. Therefore, the valley in front of the ice has been aggraded 50 feet above the part overridden by the ice. Assuming aggradation of 1 foot in 20–25 years, the time necessary to have built up this amount would be 1,000–1,250 years. Disregarding a number of factors which may modify such a computation, it is estimated that the ice advance took place 1,000–2,000 years ago. Thus the warm climate may be considered to have lasted from 5,000 or 6,000 years ago to 1,000 or 2,000 years ago, roughly from 5,500 to 1,500 years ago.

RECENT ICE ADVANCE

Evidence of the recent advance of ice is to be found in most of the valleys partially occupied by glaciers. The maximum advance noted was in Craig Valley, a tributary of the Iskut. At the mouth of this valley, 20–25 miles from glaciers which would have contributed to Craig Glacier, there is a large deposit over 100 feet high of steeply dipping, bedded materials ranging from boulder gravels to sand. The deposit could have been formed only by outwash from a glacier. Its limited extent and steeply dipping beds suggest that ice stood here for only a short period of time. As its height is about the maximum noted for any deposit of this type, probably most of the deposit is now exposed above the aggraded valley base. It is located at a point where the rock base would be expected to be at a considerable depth. Most of the aggradation from the rock base to the present level must have taken place before this deposit was formed, and therefore it is comparatively recent in age. Between this deposit and the glaciers are some recessional moraines, and even toward the heads of the valleys there are mature trees which would have required 250–300 years to grow.[7] A rough estimate of the time required

[7] Moraines known to be more than 25 years old have merely the beginning of a covering of vegetation. Alders and other deciduous trees grow first, and those where observed being replaced by evergreen trees are probably at least 25 years old. Evergreen trees on the moraines are known to be 110–50 years old.

for the building of the recessional moraines and for the growth of the trees indicates that retreat took place about 1,000 years ago. Allowing 500 years for the advance of the ice and building of the outwash hill, one arrives at a figure of 1,500 years back to the beginning of the ice advance which marked the end of the warm period.

Indian legends tell of an ice bridge over the Stikine at Great Glacier, and there are good reasons for believing that such a bridge did exist. Beyond this glacier there are two high well-defined moraines. Computing the age of the inner moraine from its rate of growth during historical times and the age of trees on the outer moraine, and adding the computed time required to build the outer moraine and to permit retreat to that position, it is estimated that the ice bridge existed 500 years ago. Anthropologists state that this is a reasonable figure. In order to have created an ice bridge, glaciers must have advanced sufficiently far to coalesce and form a glacier of considerable size in Stikine Valley, because with only a small advance Stikine River would have broken the ice away from the glacier face as fast as it advanced. Thus the ice-bridge stage of 500 years ago represents a period in the last retreat.

A period of 150 years, from 500 years ago to 350 years ago, is estimated for the retreat from the ice-bridge stage of Great Glacier to the position of the outer moraine. This was probably a period of general retreat of glaciers. Then, judging by the moraines of Great Glacier, this was followed by a halt of about 100 years, followed by more retreat (50 years), and then another halt (100 years), until about 100 years ago at the inner moraine. Until 1860 there was further retreat, and then an advance to the old position at the inner moraine. From 1910 to 1930 there was retreat to about the position of 1860. The recent history of these glaciers has thus been a succession of many minor oscillations.

EXTENT OF ICE EROSION

In the foregoing sections much information relative to the extent of ice erosion has been given but little which offers a definite clue to the maximum depth of ice erosion in any one place. As previously noted, the present channel of Stikine River is probably at an elevation not greatly different from that prior to the Quaternary. If, then,

some knowledge of the depth of the rock base of Stikine Valley could be secured, the amount of ice erosion could then be fairly accurately estimated. Valuable data can be secured because of the fact already noted that glaciers lose erosive power on debouching from a narrow valley quickly and that most valleys glaciated below sea-level are much shallower at such points than elsewhere. This is well displayed off Thomas Bay, which, though a characteristically deep fiord, is rimmed around its outer end with a shallow rock floor.

Depths in present fiords exceed 1,200 feet and are more than twice this in outer sounds and straits. Ernest Sound, just south of Stikine Strait and parallel to it, has depths of over 2,200 feet; and Behm Canal, the next channel to the south, has a known maximum of 2,328 feet. In Chatham Strait the depth reaches 2,900 feet. In Stikine Strait a maximum of only 500 feet is recorded within 3 miles of the delta. This, however, is not considered to be representative of what might be expected of the rock base, since depths of over 1,100 feet are reached at 8 miles. Certainly, it is not representative of the rock base in Stikine Valley, since Stikine Glacier, on passing from one channel less than 3 miles wide to several channels, must have lost tremendously in erosive power, so that the eroded depth outside the valley would probably be much less than inside.

Since the Stikine Valley has carried the largest volume of water and ice of any valley in this section of the range, it seems reasonable that its depths should be the greatest and should be commensurate with its great width. Therefore, the 1,200 feet now found in other fiords, which may not represent the depth to rock bottom, ought to be exceeded by a considerable amount in Stikine Valley. A projection of its walls in a profile similar to those of other glaciated valleys would easily allow a depth of 2,000 feet below sea-level in the lower reaches. It therefore seems entirely reasonable to assume a maximum depth exceeding 1,500 feet, which would also represent the amount of ice erosion. On the same basis, at the Scud, where a depth of 1,000 feet below sea-level is possible, a maximum deepening below sea-level of over 300 feet is assumed. This gives maximum erosion here of about 500 feet. The gradient thus indicated throughout seems to be more or less the normal expectation. The presence of rock islands in the valley off Porcupine River does not necessarily

indicate a shallow base, for hills over 2,000 feet high occur in Chutine Valley[8] near Pendant River.

A study of recorded depths at the outer limit of the Alaskan islands reveals that all channels in crossing the sea platform are less than 1,200 feet deep for stretches. Depths up to 2,900 feet have been recorded 90 miles east along the channels. Allowing for only a very moderate gradient and moderate cutting by ice in the channels on the sea platform, it is evident that an estimate of 2,000 feet for the maximum deepening on these data alone is plausible.

On Dease Lake in northern British Columbia the amount of deepening fairly well established is between 400 and 600 feet.[9] Here the ice was confined in a valley of relatively low banks, was moving up the slope, and probably acted over a comparatively short time. If it was capable of cutting so much under these conditions, it seems logical to expect that the great volume of ice forced through the narrow and deep valleys of the coastal areas, both within and beyond the mainland coast, would erode much more deeply.

There is no reason whatsoever for believing that because the ice base was well below sea-level the erosive power would be appreciably decreased. The buoyancy of the sea would affect only the outer limits of the glacier; and even there, unless the extent of the ice above sea-level decreased below 400 feet, the effect would be barely appreciable on a mass extending 2,600 feet (2,000 below present sea-level plus 600 feet for Pleistocene submergence) below sea-level. However, it is probable that a glacier such as that of the Stikine with a below–sea-level thickness of more than 2,100 feet would maintain an ice front of not less than 400 feet above sea-level, so that buoyancy, other than on individual fragments, may never have been a factor of much importance; certainly it could not be with ice levels resting thousands of feet above sea-level.

The figures of 1,500 feet for ice erosion within Stikine Valley and of 2,000 feet in outer straits are probably conservative. The evidence along Stikine Valley indicates that the base of the pre-Quaternary Stikine Valley was not greatly different in elevation from the

[8] *Can. Geol. Surv. Map No. 309A* (1935).

[9] W. A. Johnston, "Gold Placers of Dease Lake Area," *Can. Geol. Surv. Summary Rept.*, Part A, (1925), p. 63A.

present channel. It is possible that Stikine River then debouched into the sea at the outer limits of the Alaskan islands. In any case, if the old gradient was about the same as at present, most of the deepening below sea-level in all outer channels must be due to ice erosion. There is no definite evidence that the coastal channels are due to drowning of the coast line. They occur only in the section of the coast known to have been subjected to intense glaciation. Channels in the sea platform beyond the limits of the islands might easily have been formed by ice—in fact, such channels would be a natural expectation. The Dease Lake channel, cut in a low, relatively flat country beyond a deep trans-range valley, exemplifies this. The Stikine and other valleys show similar features on the east side of the Coast Range, where the heavily glaciated section extends some distance into the interior plateau.

STRUCTURE OF THE NORTH SANTIAM RIVER SECTION OF THE CASCADE MOUNTAINS IN OREGON

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ABSTRACT

The Cascade Mountains of Oregon in the vicinity of the North Santiam River may be divided into two structural units, the Western Cascades and the High Cascades. The Western Cascades consist of Oligocene and Miocene basaltic to rhyolitic volcanics which have been thrown into a series of northeast-southwest-trending folds, and intruded by dioritic masses with attendant mineralization. The High Cascades are composed of several distinct series of basaltic to dacitic volcanics which are Pliocene and Pleistocene in age. The young eruptives retain their initial dips and in part lap over the Western Cascades. The eastern boundary of the Western Cascades is determined by the east-facing Cascade fault scarp which obliquely cuts the folded structures. It is suggested that the High Cascades occupy a zone of weakness between the Western Cascade folding and the essentially east-west structures of eastern Oregon, and that the Cascade fault is the westernmost of a series of buried fractures of the basin range type.

INTRODUCTION

The North Santiam River region is in northwestern Oregon about 50 miles south of the Columbia River. The section studied extends 70 miles eastward from the Willamette River at Independence, a few miles south of Salem, to the summit of the Cascade Mountains between Mount Jefferson and Olallie Butte, and follows the North Santiam River. At the Willamette Valley end of the section the geology of 300 square miles was mapped in detail, and at the eastern end 250 square miles were mapped on a scale consistent with the available base. In the middle portion between Mill City and Detroit, except for the mapping of a diorite mass, only sufficient field work was done to yield data on the structure.

Brief accounts of observations in the Cascade Mountains in Oregon published toward the end of the last century were so general in character as to have little bearing on the present problems. The only detailed investigations published are the classic paper on Crater Lake by Diller and Patton,[1] Williams' description of Mount

[1] J. S. Diller and H. B. Patton, "The Geology and Petrography of Crater Lake National Park," *U.S. Geol. Surv. Prof. Paper 3* (1902).

Thielsen,[2] and reports concerning small mining districts. The latest paper dealing with the entire Cascade Mountains in Oregon is that published by Dr. Eugene Callaghan of the United States Geological Survey.[3] He pointed out that south of Mount Hood the Cascade Range can be divided longitudinally into two parts, the Western Cascades and the High Cascades, on the basis of a marked unconformity in the lava sequence. This division, which is very marked south of Mount Jefferson, greatly simplifies geological description of the range.

In the North Santiam section the Western Cascades are 50 miles wide. If the Salem Hills, topographically distinct from, but structurally a part of the range, be included, the width is 60 miles. The average elevation of the western half of the Western Cascades is less than 2,000 feet, and stripped surfaces sloping to the northwest dominate the topography. The eastern half averages somewhat over 4,500 feet in elevation, is maturely dissected, and has a maximum relief of 4,000 feet. The High Cascades, which in part overlap the eastern side of the Western Cascades, are 25 miles wide. South of Mount Jefferson the young range is essentially a linear volcanic accumulation triangular in cross section, deeply incised by glaciated valleys and characterized by dip slopes. The crest line, which averages 6,000 feet in height, is about 10 miles east of the western base of the range. North of Mount Jefferson the range widens and assumes the aspect of a plateau. The drainage divide gradually decreases from 7,000 feet to 3,400 feet at Clackamas Meadows 20 miles south of Mount Hood. Small cones are scattered over the plateau.

STRATIGRAPHY

With the exception of the diorite west of Detroit, all the rocks of this section of the Cascades are volcanic in origin. In the Willamette Valley the volcanic débris was deposited under marine conditions; all other material appears to be continental.

The almost complete lack of fossils in the rocks of the Cascade

[2] Howel Williams, "Mount Thielsen, a Dissected Cascade Volcano," *Univ. Calif. Publ. in Geol.*, Vol. XXIII, No. 6 (1933), pp. 195–214.

[3] Eugene Callaghan, "Some Features of the Volcanic Sequence in the Cascade Range in Oregon," *Trans. Amer. Geophysical Union, 14th Ann. Meeting* (1933), pp. 243–49.

Mountains in Oregon makes accurate age determinations of formations very difficult, if not impossible, with the methods now available. Marine fossils occur along the western margin of the range but have not been found more than a very few miles east of the Willamette Valley. A few new fossil leaf localities were discovered in the course of this investigation. Very fragmentary leaves occur in a gray shale with cubical fracture in the Little North Santiam River about ½ mile above its confluence with the North Santiam River. Well-preserved leaf imprints were obtained from thin-bedded tuffaceous sediments in Skunk Creek, 2½ miles south of Breitenbush Hot Springs. A third promising locality was found near the middle of the east slope of Mount Bruno at an elevation of about 4,000 feet.

The formations of the Western Cascades range in age from probable Clarno to Upper Miocene. The oldest unit, the age of which has been definitely determined, is the marine Illahe formation, which may be 2,000 or 3,000 feet thick. Its base is not exposed in the area mapped. The fine tuffaceous sandstones which constitute the bulk of this formation contain an abundance of marine invertebrate fossils. Eastward the beds grade laterally into the continental Mehama volcanics, chiefly pyroclastics. The ancient shore line crossed the western part of the Stayton basin, and swung northeast along the present margin of the Western Cascades.

The Stayton lavas unconformably overlie the Illahe and Mehama formations. They are basaltic to andesitic in composition, average 200 feet in thickness, and apparently thicken northeastward. In the Western Cascade foothills the Fern Ridge series of tuffs and conglomerates, ranging up to 1,500 feet in thickness, conformably overlie the Stayton lavas.

West of Mill City andesitic lavas are so abundant in the Mehama and Fern Ridge formations that their formational identity is lost, and the unconformity between the Mehama volcanics and Stayton lavas is not traceable. The Mehama, Stayton, and Fern Ridge formations therefore are grouped in the Sardine series, which at the type locality in Sardine Mountain northwest of Detroit consists of about 6,000 feet of basaltic to dacitic lavas. Along the Breitenbush River the Sardine series grades down into the dominantly tuffaceous

Breitenbush series some 7,500 feet thick. The basal exposed portion of the latter series consists of rhyolitic lavas and pyroclastics. Rhyolite which occurs in the western slopes of the High Cascades is considered part of the Sardine series on structural grounds.

The fauna of the Illahe formation appears to be practically identical with the Middle Oligocene assemblage at Pittsburg Bluff. Since this fauna is so distinctive, there is little doubt as to the Middle Oligocene age of the Illahe beds. The apparent intergradation of the Illahe strata and the Mehama volcanics argues strongly that the two formations are stratigraphically equivalent. The Mehama volcanics are tentatively correlated with the Eagle Creek formation of the Columbia River Gorge, because both formations are evidently Oligocene, are very similar in character, and have analogous relations to the overlying lavas.

The Stayton lavas are tentatively correlated with the Columbia River basalts. The flows appear to be of the fissure type and some can be followed individually for several miles. The lavas thicken northeastward toward the Clackamas River, where a thick section of the basalts is exposed. The magnitude of the unconformity with the Illahe beds suggests that the lavas are Miocene rather than Oligocene. The Fern Ridge tuffs are also considered as Miocene, since they apparently lie conformably on the Stayton lavas.

The Sardine and Breitenbush series may well range in age from Clarno (Oligocene-Eocene) to Middle or Upper Miocene. It is extremely likely that the Sardine lavas, although ranging from basalts to dacites, are equivalent at least in part to the Columbia River basalts. The latter have been traced,[4] with notable thickness, to the southern edge of the Estacada quadrangle, about 20 miles northeast of Sardine Mountain. Local variation in composition is one of the outstanding characteristics of Western Cascade lava series. Interfingering of more acid flows in the Columbia River basalts doubtless accounts for the petrologic difference between the two series in this district. Exactly such a lateral change was traced between basalt flows of the Stayton lavas and the Sardine series. The Sardine lavas, moreover, were mineralized by the Halls diorite.

[4] F. F. Barnes and J. W. Butler, unpublished Master's thesis, University of Oregon, 1930.

This intrusion is petrologically very similar to intrusive bodies in the Washington Cascades, which have been considered Upper Miocene.[5] The flora from the Skunk Creek and Mount Bruno localities has been identified as "typically upper Miocene."[6] Since the localities are far east of the Breitenbush anticlinal axis, the beds are stratigraphically high in the tuff series, and may be equivalent in age to some of the Sardine lavas. The rhyolitic lavas at the base of the exposed section of the Breitenbush series are tentatively correlated with the upper Clarno on the basis of lithology. It is suggested that the bulk of the Breitenbush tuffs is equivalent to the John Day series. Failure to find unconformities at the top of the rhyolite and at the base of the Sardine series may be due in part to lack of good exposures.

It is apparent that the Western Cascade section compares closely with that of eastern Oregon. The Sardine lava series is the presumed equivalent of the Miocene Columbia River lavas; the Breitenbush tuff series is the probable equivalent of the Oligocene John Day tuffs, the basal rhyolite members being the correlative of the rhyolite in the upper part of the Clarno. The diorite west of Detroit is probably Upper Miocene, since it was intruded into the Sardine series.

The unconformity between the Western Cascade eruptives and the High Cascade lavas is marked. The older rocks had been folded, mineralized, and intruded by dioritic plugs, probably lifted as a block by faulting, and deeply eroded before the earliest of the High Cascade rocks were poured out. This unconformity is recognizable the entire length of the Cascade Mountains in Oregon.

The High Cascade lavas in this district are chiefly light gray, porous, fine-grained types. They range from olivine-bearing basalts to andesites containing notable amounts of cristobalite. The basaltic types predominate, some containing both olivine and cristobalite. The lavas are divided into four groups, of which only the youngest, that forming the modern peaks, is composed mainly of andesites. In chronological order these groups are: the Outerson basalts, ranging up to 3,000 feet in thickness; the Minto basalts, attaining a prob-

[5] G. O. Smith and Bailey Willis, "Contributions to the Geology of Washington," *U.S. Geol. Surv. Prof. Paper 19* (1903), p. 22.

[6] Ralph Chaney, personal communication.

able thickness of 4,000 feet and forming the bulk of the High Cascades; the Santiam basalts, filling the North Santiam River valley to a depth of 1,800 feet; and the Olallie lavas of the modern volcanic peaks. Since the two last-named groups are nowhere in contact, their relative ages are unknown; similar degrees of erosion suggest that they may have been nearly contemporaneous. The Battle Ax basalts which were poured out from a source near Battle Ax Mountain, northeast of Detroit, are tentatively correlated with the Minto lavas on the basis of very close petrographic similarity and comparative amount of dissection of the flows. At the present time it is only possible to assign the High Cascade lavas to the Pliocene and Pleistocene. The Outerson basalts rest with marked unconformity on the tuffaceous beds in the upper part of the Breitenbush series, which have been dated as Middle or Upper Miocene by means of plant remains. The Santiam basalts are pre-Wisconsin, but undoubtedly Pleistocene.

STRUCTURE OF THE WESTERN CASCADES

As outlined above, the Western Cascades consist of a thick mass of volcanic rocks of various types. These rocks have been compressed into a series of gentle folds trending northeast-southwest, the folding being more intense toward the southeast. The folds, in sequence from west to east, are the Willamette syncline, Mehama anticline, Sardine syncline, and Breitenbush anticline (Fig. 1). All apparently plunge toward the northeast. The eastern limit of the Western Cascades is probably determined by the Cascade fault which cuts off the eastern limb of the Breitenbush anticline.

The Willamette syncline and Mehama anticline are very broad open folds. Dips in the Miocene lavas average less than 3°; in the Oligocene formations dips approximate 10°–12° and in one place reach verticality. The western limb of the syncline is formed by the Salem Hills homocline. The synclinal axis trends about N. 30° E. and plunges northeastward at a low angle. The nose of the fold, as indicated by the Stayton lavas, lies east of the south end of the Salem Hills. The low Turner cross-warp southeast of Salem separates the Stayton basin from the main syncline. The closure of the Mehama anticline in the Stayton lavas is about 1,000 feet, the

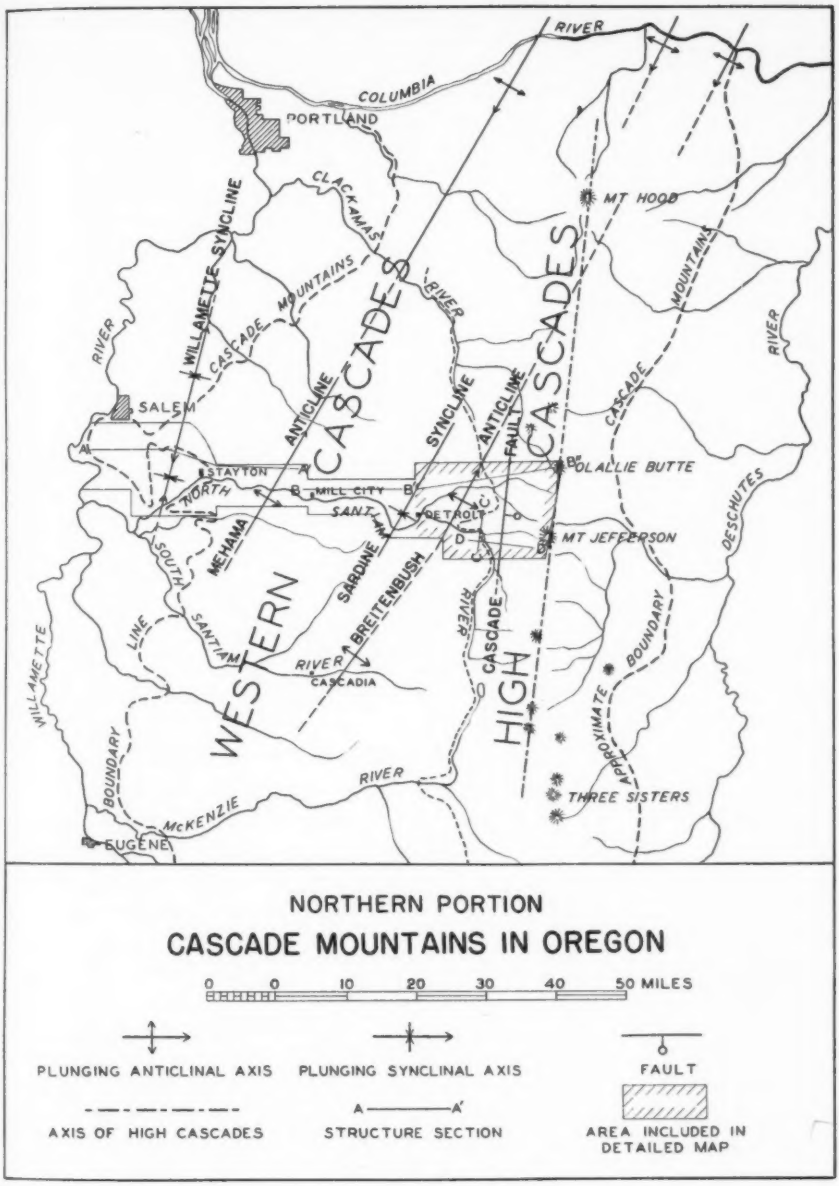


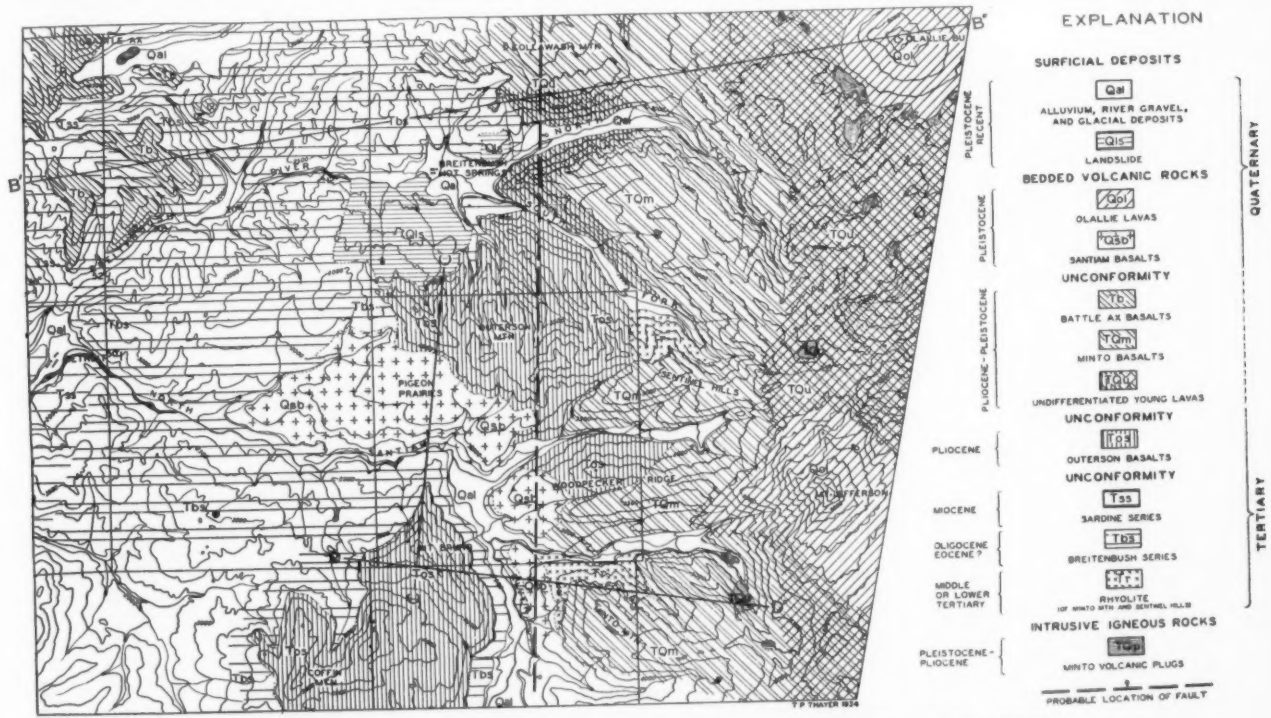
FIG. 1.—Northern portion of the Cascade Mountains in Oregon

anticlinal axis lying about 6 miles west of Mill City. The dip slopes which characterize the western margin of the Western Cascades are formed on the western limb of this structure. The eastern limb of the fold descends into the Sardine syncline 3 miles west of Detroit.

This syncline and the Breitenbush anticline are very asymmetrical structures. The beds in the western limb of the Breitenbush anticline dip 50° westward, while those in the eastern limb dip 12° eastward. Northward plunge of the structure is indicated by northerly dips north of the Breitenbush River, and by greater width of the rhyolite exposures along the axis to the southwest. The angle of plunge appears to be about 10°. The crest of the fold is marked by minor faults and fracture zones. Intrusion of porphyritic diorite occurred in the axis of the Sardine syncline. The intrusive body, about one mile across, left the lavas practically undisturbed, suggesting that intrusion was by stoping or assimilation rather than by forcible injection. The volcanic neck in Battle Ax Mountain is also situated near the synclinal axis.

THE CASCADE FAULT

The eastern margin of the Western Cascades is the eastward-facing buried Cascade scarp which trends N. 10° E. It extends south from Collawash Mountain, northeast of Breitenbush Hot Springs, beneath Outerson Mountain, and through the eastern side of Mount Bruno (Fig. 2). The slope of the scarp, as revealed by the basal contact of the Pliocene lavas, varies from 1,500 feet per mile in Collawash Mountain to an average of 500 feet per mile in Mount Bruno. The height of the scarp was at least 2,000 feet at the time of burial by the High Cascade lavas. Careful examination of the western slopes of the High Cascades revealed only two exposures of older rocks buried under the Minto lavas, namely, the rhyolite in Minto Mountain and northwest of the Sentinel Hills. The rhyolites dip about 15° eastward in both localities, and appear to overlie andesitic rocks. In Minto Mountain they attain an elevation of 4,250 feet, and in the Sentinel Hills slightly over 5,000 feet. An intrusive plug in the Sentinel Hills suggests that the rhyolite may have been locally erupted. These lavas are the only pre-High Cascade rocks found east of the Cascade scarp.



GEOLOGY OF PORTIONS OF THE MILL CITY AND MT. JEFFERSON QUADRANGLES, OREGON.

0 1 2 3 4 5 MILES

CONTOUR INTERVAL 500 FEET

FIG. 2.—Geology of portions of the Mill City and Mount Jefferson quadrangles, Oregon

The character of the Cascade scarp and its structural relations suggest faulting. Physiographic evidence indicates that the scarp, although probably reduced in height and steepness, was quite straight at the time of burial, in spite of the fact that it cuts across rocks differing very greatly in resistance to erosion. Moreover, the scarp cuts diagonally across the strike of the beds, which dip toward the face. In brief, the structural conditions are practically the opposite of those required for the formation of an escarpment by erosion.

Two explanations for the distribution of the pre-High Cascade rocks and the buried Cascade scarp immediately come to mind. The simpler is that the rhyolite masses are outliers of the Western Cascades and that the scarp is an erosional feature. In view of the structural relations outlined above and the fact that the erosion must have taken place in the lee of the range, this hypothesis is considered untenable.

A second hypothesis is that the scarp is the result of faulting and that the buried rhyolite formed mountainous masses on the hanging-wall block. This accounts for the straightness of the scarp and the scarcity of old rocks to the east on the dropped side of the fault. The age of the rhyolite is a question on which this hypothesis may stand or fall.

The present elevation of the rhyolites indicates that they must have stood considerably above the main Western Cascade block before faulting occurred. If of Clarno age, discordance of structure in the rhyolite and Western Cascade rocks is to be expected, as the unconformity between the Eocene and Oligocene is markedly angular elsewhere; disconformable relations would be a coincidence. Also, if the rhyolite is Clarno, it must be the stratigraphic equivalent of the rhyolite exposed in the axis of the Breitenbush anticline, if it is not older. If such is the case, the thickness of Clarno is at least 7,000 feet. This figure is based on the conservative assumption that the average eastward dip is 10° and that no faulting has occurred. If the dip is taken as 12° and a throw of 2,000 feet be postulated for the fault, the thickness of Eocene is 10,000 feet, a very unlikely figure. Furthermore, the base of such a high mountain mass should have extended some distance to the west and should be found in the scarp.

The absence of rhyolite in Mount Bruno and the North Santiam River west of Minto Mountain proves that such was not the case. It seems very probable, therefore, that the rhyolite east of the Cascade scarp is part of the Oligocene-Miocene sequence and equivalent to the later rhyolitic lavas of the Sardine series.

According to this hypothesis the Cascade scarp was formed by a fault having a throw of at least 2,000 feet, the western side being raised. The rhyolite hills buried by the High Cascade lavas are regarded as high-standing masses built up by local eruptions on the eastern border of the Western Cascade range, which later was faulted down. The presence of the greatest eruptive center of the Outerson basalts at the foot of the Cascade scarp where the fault presumably passes, is rather suggestive of volcanic response to fracturing.

STRUCTURE OF THE HIGH CASCADES

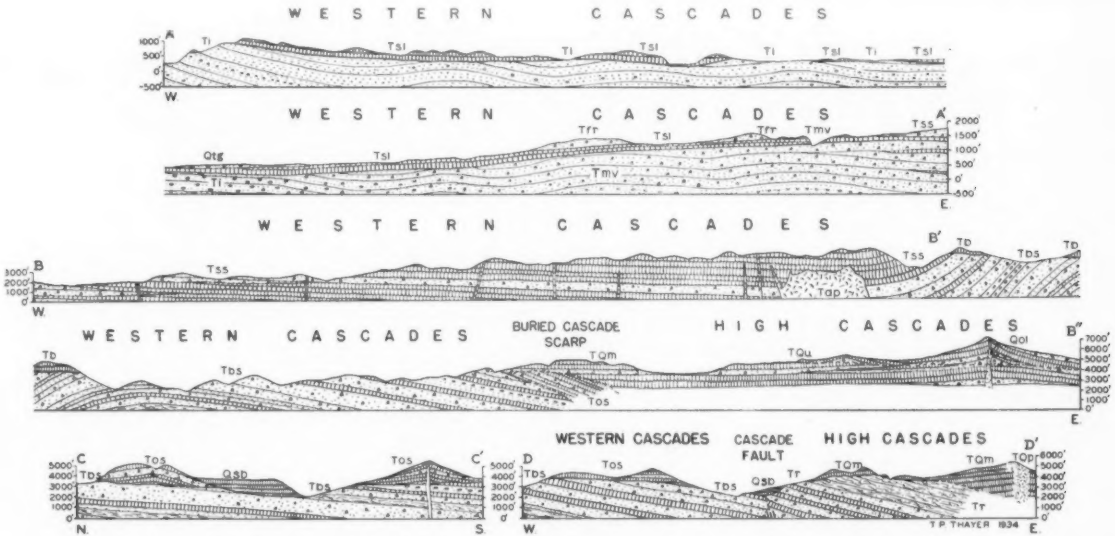
As stated above, the High Cascade lavas differ markedly from the Western Cascade series in that they retain their initial dips. The earliest series, the Outerson basalts, was extruded from sources along the Cascade fault and on the scarp. The greatest center of eruption in this district was located east of Outerson Mountain (Fig. 2). Other centers were scattered along the scarp from Mount Bruno southward. The Outerson lavas and pyroclastics, having been erupted on the Cascade scarp, in general dip eastward at angles varying from 1° to 20°. That these basalts flowed westward beyond the scarp onto the general upland of the Western Cascades is indicated by the flatness of their basal contact in the vicinity of Coffin Mountain. The Minto basalts, which form the bulk of the High Cascades, were extravasated from volcanic plugs of cristobalite-bearing micronoritic type on or near the present crest of the range. The dips in these lavas vary from 7° or 8° near the plugs to 1° or 2° at the margin of the range, as in Collawash Mountain. These low dips contrast strongly with the high dips, as much as 28°, in the Olallie lavas which make up the young peaks of the present time. The Park Butte plug north of Mount Jefferson was near the center of a huge shield-type cone whose slopes extended northwestward beyond Collawash Mountain and northward past Olallie Butte. From Outerson Mountain north the Minto lavas completely buried the

Outerson eruptives and the Western Cascade margin; as a result there is no physiographic boundary between the two ranges. South of Outerson Mountain the later lavas flowed around the Outerson accumulations, only partially burying them. The course of the North Santiam River was in all probability determined by the intersection of the Minto initial surface with the eroded surface of the east-dipping Outerson basalts. This trough, now greatly modified, is the physiographic boundary between the Western and the High Cascades from Outerson Mountain to the McKenzie River west of the Three Sisters. Probably contemporaneously with the eruption of the Minto basalts the Battle Ax basalts were poured out on the Western Cascades north of Detroit (Fig. 2). The topography at the time of eruption was mountainous, with a relief of at least 3,000 feet. Battle Ax Mountain appears to have been the southernmost peak of a chain of volcanoes which extended northeastward toward Mount Hood.

The Santiam basalts partially filled a broad valley cut in all the older rocks. They were poured out from an unknown source, probably of the fissure type, near Outerson Mountain and flowed westward and southward from the elbow in the North Santiam River. At the present time these lavas extend halfway to Detroit from the elbow. Originally they must have extended much farther west. They flowed southward beyond the southwestern corner of Minto Mountain. The width of the valley which these lavas filled indicates that the Santiam basalts are considerably younger than the Minto lavas. The unconformable relations of the Olallie andesitic lavas to the Minto basalts, shown in Section B'–B'' of Figure 3, indicate a comparable age difference, suggesting contemporaneity of the two youngest series.

REGIONAL STRUCTURAL RELATIONS

The Western Cascade structures in the North Santiam section and the Columbia River section are very similar and are in alinement. The northeastward projection of the Mehama anticline approximately coincides with the Eagle Creek anticline of the Columbia River gorge; both folds are very broad and gentle. The projection



GEOLOGIC STRUCTURE OF THE NORTH SANTIAM RIVER SECTION
OF THE
CASCADE MOUNTAINS IN OREGON

0 1 2 3 4 5 6 7 8 9 10 MILES

Qtg	TERRACE GRAVEL	TQp	MINTO VOLCANIC PLUGS	Ti	ILLAHE FORMATION
Qol	OLALLIE LAVAS	Tb	BATTLE AX BASALTS	Tmv	MEHAMA VOLCANICS
Qsb	SANTIAM BASALTS	Tos	OUTERSON BASALTS	Tss	SARDINE SERIES
TQu	UNDIFFERENTIATED YOUNG LAVAS	Tfr	FERN RIDGE TUFF	Tbs	BREITENBUSH SERIES
TQm	MINTO BASALTS	Tsl	STAYTON LAVAS	Tr	RHYOLITE
		Tdp	DIORITE PORPHYRY		

FIG. 3.—Geologic structure of the North Santiam River section, Oregon

of the Breitenbush anticline lies in the vicinity of the Ortley and Bingen anticlines. The two last-named folds are sharp but smaller than the Breitenbush structure, and taken together may be its equivalent. The southwestward projection of the Breitenbush anticline crosses the South Santiam River not far from the steep structures east of Cascadia mentioned by Hodge.[7] This structural alinement of the three sections strongly suggests that the folds of the Western Cascades are persistent for long distances and cut diagonally across the main north-south trend of the mountains.

Whereas the folds trend northeast-southwest, the veins and associated andesitic and dioritic dikes of the entire Western Cascade system in Oregon strike northwest. This discordance of trends, the plunging character of the main folds, and the Turner warp (p. 706) imply that the intrusion and mineralization followed minor warping and fracturing along northwest-southeast axes. The contact relations of the Halls diorite, as stated above, indicate emplacement by stoping and assimilation; and that adjustment of compressive forces was nearly, if not quite, complete before intrusion occurred. As the intruded rocks are believed to be Middle Miocene in age, the folding and intrusion are thought to have taken place during Upper Miocene time, probably toward the end of the Miocene period, simultaneously with the Miocene orogeny in the Washington Cascades.[8] The present elevation of the Western Cascades is considered to be almost entirely due to the uplift which occurred in Pliocene time when the region west of the Cascade fault was raised and possibly tilted toward the west, probably as a more or less rigid block. The faulting probably occurred not later than the Middle Pliocene. Evidence of elevation by arching, as described by Willis and Smith[9] is very obscure, if not absent, in this portion of the Cascade Mountains.

Pre-Pliocene folds of eastern Oregon, such as the Ochoco Mountains, Tygh Ridge, and the Mutton Mountains, in general trend

[7] E. T. Hodge, "Framework of the Cascade Mountains in Oregon," *Pan-Amer. Geol.*, Vol. XLIX, No. 5 (1928), p. 346.

[8] Smith and Willis, *op. cit.*

[9] *Ibid.*, pp. 39, 85–86.

east-west. These structures disappear under the eastern slope of the High Cascades but have not been recognized in the Western Cascades. Folding in the Columbia River basalts probably occurred at the same time as that in the Western Cascades, since the Pliocene volcanics of eastern Oregon are undisturbed. Minor cross folding in the Western Cascades has been mentioned, but the major adjustment between the two systems of folds appears to have been accomplished in the area now buried under the High Cascades. This area seems to be a weak zone in which faulting may have occurred during and after the Miocene folding.

Faulting of the basin and range type has been described in the Klamath Lake region by Gilbert,[10] and the scarp south of Crater Lake is commonly regarded as a fault scarp. It does not seem improbable that faulting of this type should occur farther north and be buried under the High Cascade lavas. Close relation of faulting and volcanic activity has been noted in many parts of the world. The linear distribution of the High Cascade volcanic centers is suggestive of such a relation. The main peaks of post-Minto age, as well as the Minto plugs, are restricted to a narrow belt along the crest of the High Cascades. From Olallie Butte south this belt is almost a line, which is parallel to the Cascade fault. Toward the north the distribution of eruptive centers is more haphazard and the High Cascades widen to form an irregular plateau. Superposition of this High Cascade plateau on the Western Cascades in the Mount Hood region is caused by intersection of the structural axes of the two ranges.

In view of the foregoing facts, it seems reasonable that the distribution of High Cascade vents should be controlled by a nearly north-south fracture system. Weakening of the fractures toward the north, evidenced by their paucity in the Columbia River gorge and abundance in the Klamath Lake region, would account for the change in areal relations of the young volcanic cones and plugs. The faults may have been an important mode of adjustment between

[10] G. K. Gilbert, "Studies of Basin Range Structure," *U.S. Geol. Surv. Prof. Paper 153* (1928), pp. 76–85.

the Western Cascade structures and those in eastern Oregon. The probable existence of a series of related fractures rather than a single fault suggests the possibility that the faulting is of the basin range type.

ACKNOWLEDGMENTS.—The writer wishes to express his appreciation of the advice and help given by Dr. J. P. Buwalda and Dr. Ian Campbell, of the California Institute of Technology. Gratitude is also due Dr. Eugene Callaghan, of the United States Geological Survey, for his suggestions and information so generously given during the course of the investigation.

NOTES ON THE JOINTING OF THE DEVIL'S TOWER, WYOMING

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ABSTRACT

Observations have shown three joint systems other than the columnar system for which the Devil's Tower is famous. These systems are conspicuous in the basal cliffs where the joints are arranged in peripheral, radial, and up-curving groups.

It is believed that these joint systems indicate an *original* shaftlike mass which has been only slightly modified to give talus accumulations.

Similarity of jointing in one group of volcanic necks and its absence in known laccoliths seem to favor the former origin of the Devil's Tower.

INTRODUCTION

The Devil's Tower in Wyoming, at the northwest end of the Black Hills, is widely known as an outstanding example of columnar jointing and also as a remarkable physiographic feature. More intimate acquaintance with this unusual mass, however, brings out the fact that all of the jointing is not of a columnar type and that probably the portions which show other than columnar jointing are important in a theory of origin. Careful reading of the detailed descriptions by Jaggar[1] and by Darton[2] brings out many of the variations in jointing. It is to emphasize and to extend knowledge of these variations, which are usually neglected, that the following descriptions are offered. The data have been gathered during annual visits with University of Minnesota geology field classes extending over several years.

SUMMARY OF PREVIOUS OBSERVATIONS

The earliest detailed descriptions of the Devil's Tower were given by Henry Newton and Walter P. Jenney.[3] Newton visited the tower in 1875 and described it in some detail on pages 201 and 202 of the

[1] T. A. Jaggar, "The Laccoliths of the Black Hills," *U.S. Geol. Surv., 21st Ann. Rept.*, Part III (1900), pp. 163–303.

[2] N. H. Darton, "Geology and Water Resources of the Northern Portion of the Black Hills," *U.S. Geol. Surv. Prof. Paper 65* (1909).

[3] "Report on the Geology and Resources of the Black Hills of Dakota," *U.S. Geog. and Geol. Surv. of the Rocky Mountain Region* (Washington, 1880).

report cited, and Jenney also mentioned it briefly on page 284. Jaggar[4] repeated most of Jenney's and Newton's descriptions, making them more accessible to those interested.

It should be emphasized that the rock of which the tower is composed is not basalt, as has been stated in at least one book, but is a coarse phonolite porphyry. A complete analysis, as well as a petrographic description by Johannsen, is given in the "Devil's Tower Folio"[5] of the United States Geological Survey.

The most complete descriptions, with theories of origin, are given by Jaggar[6] and by Darton and O'Harra.[7] The present writers have drawn freely on these descriptions as well as on their own observations.

DESCRIPTION OF THE TOWER

General.—As shown by published descriptions and photographs, the tower is not exclusively a columnar mass, but consists of two main divisions, a broad basal portion above which rises a smaller columnar mass. For a study of the joint structure it may be somewhat further subdivided and described under the following heads: base, shoulder, main columnar portion, crest, and finally the talus.

A contour map recently made by the United States National Park Service gives the elevation of the highest point as 5,117 feet. Although the Belle Fourche River flows close to the base on the east side at an elevation of 3,835 feet, the upland has an average elevation of 4,325 feet. The tower proper therefore rises above its immediate base about 800 feet.

The rock which comprises the tower is a comparatively uniform phonolite porphyry. Examination of the talus blocks, as well as the accessible portions of the rock in place, reveals little variation in texture. The largest phenocryst found by the writers measured $\frac{9}{10}$ inch in length, but the average length is about $\frac{1}{2}$ inch. Thin sections of the porphyry show a groundmass of fine trachytic texture. Because of the porphyritic texture, a study of oriented thin sections

[4] *Op. cit.*

[5] N. H. Darton and C. C. O'Harra, "Devil's Tower Folio," *U.S. Geol. Surv. Folio 150* (1907).

[6] *Op. cit.*

[7] *Op. cit.*; Darton, *op. cit.*

might reveal information as to the direction of movement of the magma which formed the mass.

As is well shown in photographs published by Darton and by Jaggar, the tower is not symmetrical but presents a decidedly differ-

FIG. 1.—Near view of northwest side of the Devil's Tower showing the remarkable regularity of the jointing.

ent outline and arrangement of columns according to the side from which it is viewed. The most symmetrical outline, as well as best-developed jointing, is seen when viewed from the northwest side preferably somewhat west-northwest (Fig. 1). From here the inclination of the sides *appears* to be about 75°. Viewed from other locations, the north side is apparently more nearly vertical, whereas

the west and east sides are inclined only 65°–70°. These are apparent dips which are flattened somewhat by the angle at which the mass as a whole must be viewed. On the south side, the actual angle of inclination in the columnar portion was measured as 85°.

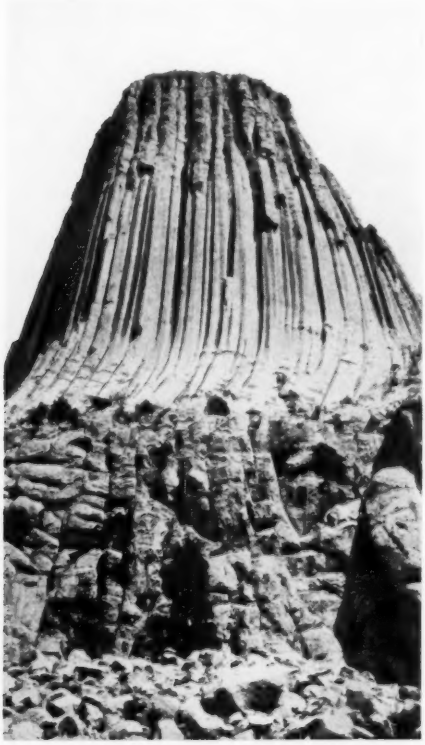

FIG. 2.—View of the southwest side of the Devil's Tower showing the conspicuous difference between the upper and lower portions. Note particularly the block jointing in the lower portion.

Base.—The base is roughly circular in outline, but is slightly elongated in a direction N. 30° E. The diameter in this direction between the bases of steep cliffs is approximately 880 feet. The minimum diameter is about 700 feet, and the average roughly 800

feet. This basal portion presents high peripheral cliffs, but locally the talus extends up over the greater part of the base.

The basal portion has jointing distinctly different from that of the columnar part, as is well shown in Figure 2. On the southwest side of the base, there is a particularly massive vertical cliff with little talus. At this locality there is no evidence of columnar structure; but large vertical, inclined, and horizontal joints divide the cliff into major blocks whose greatest dimensions are 10–12 feet. This basal portion passes rather gradually into the higher columnar portion, without a sharp boundary between the two parts.

TABLE I

OBSERVATIONS ON JOINTS OF THE DEVIL'S TOWER

Location	Trend of Cliff	Peripheral Joints		Radial Joints	
		Strike	Dip	Strike	Dip
N.NW. W.NW.	NE.–SW.	N. 65° E. N. 40° E.	67° NW. 90°	N. 60° W.	90°
W.	No cliff; talus cover				
SW.	NW.–SE.	N. 30° W.		N. 50° E.	35° NW.
S.	E.–W.	N. 90° E.	75°–80° S.	N.	90°
SE.	NE.–SW.	N. 45° E.	78° S.	N. 45° W.	50° SW.
E.	N.–S.	N.	80° E.	N. 80° E.	48° N.
NE.	NW.–SE.	N. 45° W.	80° E.	N. 45° E.	90°
N.	E.–W.	N. 90° E.	75°–80° N.	N.	90°

The data on joint systems in the cliffs of the basal portion are shown in Table I. These observations show that there are two important joint systems, namely, peripheral and radial. The cliffs of the base are along large vertical joints which are arranged in such a manner as to more or less parallel the periphery of the tower. The radial system is prominently represented by those joints which are perpendicular to the peripheral cliffs.

Shoulder.—The shoulder is in reality the transition from the cliffs of the wide basal portion to the main columnar part of the tower (Fig. 3). In many respects this is the most interesting part of the entire structure, perhaps the more so because it is possible to climb over it for detailed observations. The configuration of this portion

is apparently related to broad, gently upward-curving joints which divide the rock in tabular masses. The prominent joints of this system seem to merge into the "columns" of the tower, and it does not appear possible to determine a definite lower limit to the columnar structure.

The base and shoulder therefore contain three systems of joints. The peripheral system determines the cliffs. The upward-curving system determines the shoulder. The radial and the upward-curving systems seem closely related to the prominent columnar jointing of the tower.

FIG. 3.—Shoulder on northwest side looking south 35° west

On the east side the shoulder largely disappears; and the columns rise parallel to the radius instead of following a slightly spiral ascent, which is more common. Since at this place the inner portion of the tower is more exposed, the possibility is suggested that the columnar portion extends farther down in the *interior* of the base. It is further suggested that, where the basal cliffs are well developed, the present configuration of the tower may not differ greatly from the original shape of the intrusion.

Main columnar portion.—The columnar portion has been described so completely by Jaggar and by Darton that it is unnecessary to repeat the full descriptions. The following points, however, seem

worthy of further emphasis. The more massive basal portion passes gradually into the curving columns which, from a dip as low as 16°, rise in broad sweeping curves to a nearly vertical attitude on the west side. At places the columns curve somewhat along the periphery as they rise. At other places, particularly on the south and northwest sides, the columns appear to rise along the direction of the radius with a minimum dip of 37°. The first evidences of the columnar structure are broad joint surfaces curving upward with transverse ribs and joints. The columns of the upper portion are not all six-sided, but some are four- and others five-sided. Darton gives the average thickness as 6 feet, but they vary up to 18 feet. Many of the columns vary in size somewhat along their length, with a general tendency to become smaller at the top, as shown by the general shape of the mass. High up in the mass some large columns are observed to divide into two columns above—probably a third one being hidden by the others. The large columns usually swell at the place where they divide into the two or more columns.

The extreme regularity of the columns when viewed at a distance is only apparent, as actually they are always somewhat irregular. Some columns are enlarged along their length, and the adjacent ones are smaller to equalize the total volume. At places the lower portions of the columns are cut by rather inconspicuous block joints. The elevation at which the columns start varies, as noted above, with the distance from the center of the mass. Where more columns have fallen away, those thus exposed extend considerably lower than those which were out over the shoulder. This difference is perhaps as much as 100 feet.

Crest.—The upper 100 feet of the tower present a decidedly ragged or rough appearance because the beautifully symmetrical columns are broken and irregular (Figs. 1 and 2). Study of this portion with powerful binoculars, as well as of its broken blocks on the talus (Fig. 4), shows that the rough appearance is due to the fact that an irregular blocky jointing has been superimposed on the columnar jointing. These joints are spaced from 8 to 18 inches apart, and on the talus fairly spheroidal masses may be seen. Weathering has affected the rock along these joints, thus emphasizing them and causing the rock to become somewhat crumbly. This

is in reality spheroidal weathering, although in most cases the process has not developed the distinctly spheroidal forms in abundance; however, enough are visible to clearly demonstrate the tendency. It may be well to state that this weathered upper portion does not seem to have any different composition or texture from the main mass.

The distinctiveness of the weathered upper portion suggests that it has been exposed to weathering much longer than the lower portion; perhaps an old erosion surface surrounded the upper portion, so that it was subjected to weathering for a long time before the lower portion was uncovered.

FIG. 4.—Talus block from upper part of tower showing tendency to spheroidal weathering.

Such a period of weathering might be correlated with the construction of the Tertiary piedmont alluvial plain. To the northwest the Big Horn River has been considered to have become superimposed upon the northern end of the Big Horn Range during the early Tertiary and has subsequently eroded a gorge approximately 2,000 feet deep. The summit of the tower is 1,342 feet above the Belle Fourche River at its base, and thus the weathering of the top might be related to these Tertiary events; but stream rejuvenation has re-

sulted in uncovering the remainder of the tower too recently to be weathered.

Talus.—The talus slope, particularly on the northwest side, is a conspicuous feature of the mass. The size of the columns is most easily appreciated by climbing over the talus (Fig. 5). The largest talus block measured was found on the northeast side. It is 53 feet long, 18 feet across the end, with each face of the hexagonal outline about 10 feet wide. The estimated weight is over 1,150 tons.

FIG. 5.—View of talus slope on west side

The talus is by no means evenly distributed around the tower. Where the basal cliffs rise highest, there is relatively little talus; but elsewhere, notably on the west side, the talus extends well up the basal portion, and the upper portion of the tower is somewhat indented as a result of many columns having fallen to form the talus. In general, talus seems to extend out about 800 feet from the basal outcrops and forms an upland from which the tower rises abruptly. Probably at no place do the talus blocks extend out more than 1,200 feet from the base, although small pieces have been carried out much farther by normal transportational processes.

DISCUSSION

The foregoing descriptions of the Devil's Tower show the numerous types of jointing involved. It is doubtful if a single relatively

small mass anywhere shows a more complicated series of fractures. The origin of the tower, particularly its original form, has been debated; and there is apparently still a lack of unanimity of opinion. It is clearly related to the series of Tertiary laccoliths and subordinate dikes and sills which occur over a large area in the northern Black Hills. For an excellent areal map showing the relations of the entire group the reader should refer to Plate IV of *United States Geological Survey Professional Paper 65*. This association, together with the exposure of nearly horizontal sediments around the base of the tower, but generally at a lower level and dipping somewhat in toward the mass, has logically led to the opinion that the mass is the remains of a laccolith.

Jaggar[8] suggested that the tower was but a remnant of a large laccolith which originally extended to Missouri Buttes, four peaks composed of phonolite porphyry, 3 miles to the northwest. Darton and O'Harra[9] suggested that the tower was a laccolith, restricted in its original areal extent; but they assumed that considerable erosion had removed the evidence of its original form. Russell[10] believed that the tower represents an intrusive mass of greater vertical than horizontal extent, and called the form a "plutonic plug"; but he also included under the term those Black Hills occurrences designated as "laccoliths" by other geologists.

In the consideration of the facts developed during the present study of the jointing the writers inevitably sought for the explanation of the variations in fracture patterns. Perhaps foremost was a possible explanation of the series of radial and peripheral joints which are well exposed in the cliffs of the base. It was also noted that this jointing did not necessarily continue in to the center of the tower. If these joints do not continue inward, they must be related to cooling of the outer portion, which in turn suggests that the cliffs facing outward from the portion of the igneous mass farthest from

[8] *Op. cit.*, p. 264.

[9] *Op. cit.*, p. 5.

[10] I. C. Russell, "Igneous Intrusions in the Black Hills," *Jour. Geol.*, Vol. IV (1906), pp. 23–43.

the center represent the approximate outer limit of the original rock mass. This seems supported by the fact that the rather straight columnar portion of the mass extends considerably farther down in the interior of the mass, as shown at the places where erosion has worked farthest into the tower. The talus slopes at present seem to check roughly with the amount of material which would be furnished by the erosion of an *original* tower-like mass. It is also evident that most of the present talus has come from those upper portions which are unprotected by basal cliffs.

It is the writers' suggestion, therefore, that the tower represents an uncovered shaftlike mass, the present base of which is indicative of its original size. Its general shape has probably been *only slightly modified* by the falling of sufficient columns to produce the talus and result in the conspicuous subdivisions of base and columnar parts.

The results of the present study concerned largely the original form, but further evidence as to the probable origin of the mass was also recognized. In his study of the volcanic necks in the Mount Taylor region of New Mexico, Johnson[11] observed and described several occurrences of columnar jointing which may be very similar to that in the Devil's Tower. Furthermore, he states that the sedimental layers remain in essentially horizontal attitudes to the very contact with the igenous material. Bonney[12] states that horizontal or inward-dipping strata are commonly present at the boundary of volcanic necks on the Fifeshire coast of Scotland. He also states that in the volcanic necks at St. Andrews, the "Spindle" is a cylindrical mass composed of radiating columns. Although the jointing of the Inyankara laccolith appears to be of the columnar pattern in Darton's[13] photograph, actual examination proved that the "columns" are the result of intersecting tabular joints rather than true columnar ones. Such descriptions as could be found of jointing in lacco-

[11] D. W. Johnson, "Volcanic Necks of Mount Taylor Region," *Geol. Soc. Amer. Bull. 18* (1907), pp. 303–24.

[12] T. G. Bonney, *Volcanoes* (1899), pp. 176, 180.

[13] N. H. Darton, "Sundance Folio," *U.S. Geol. Surv. Folio 129* (1905), Fig. 5.

liths mostly refer to the platy or slabby character which results from the intersecting systems of jointing.[14]

Since the jointing of the Devil's Tower is similar to that of at least one group of undoubted volcanic necks and since no description of that pattern of jointing in a laccolithic mass is known to the writers, it seems more logical that the Devil's Tower should be considered as a volcanic neck until more evidence is found for the alternative occurrence by new structures in the Devil's Tower or by similar jointing in unquestionable laccoliths.

[14] Some "columnar" jointing has been described as occurring in the laccoliths of the Judith, Highwood, and Little Belt mountains of Montana. In some cases the evidence for such designation of the intrusive form is not unquestionable, and none of them show the basal flare which is characteristic of the columns in the Devil's Tower and the volcanic necks of the Mount Taylor region.

INERTIA AS A POSSIBLE FACTOR IN THE MECHANICS OF LOW-ANGLE THRUST FAULTING

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ABSTRACT

This paper attempts to evaluate the importance of inertia in the mechanics of the large low-angle thrust faults. Because of a lack of precise information many assumptions have been necessary, involving both the mechanics of thrusting and the numerical values of the factors involved. The analysis indicates that under certain hypothetical conditions inertia may be a factor of great importance. Further field data will have to be obtained before it can be determined to what extent these principles can be applied in nature.

The problem of the mechanics of the large low-angle thrusts was brought to the writer's attention through a study of the Heart Mountain thrust near Cody, Wyoming. Several peculiarities of this fault have been noted by W. H. Bucher,[1] one of which is that the section of strong rocks in the thrust plate is too thin (1,500 feet) to transmit powerful stresses over long distances. In calculating the theoretically possible maximum displacement of low-angle thrusts of this type, the writer reached the same conclusion that he later found had been reached by Lawson[2] in 1922, namely, that the rocks in any of the large low-angle thrust plates are not strong enough to transmit the force necessary to move this plate against the friction on the thrust plane. While discussing this point with Dr. Chamberlin, he suggested an investigation of the possible importance of inertia as a heretofore neglected factor in the mechanics of thrust faulting.

In the following consideration of the mechanics of low-angle thrust faults the writer applies the terms "overthrust" and "underthrust" only to low-angle thrusts of considerable displacement, and uses the earth's center of gravity as a point of reference. Thus the term

[1] W. H. Bucher, "Volcanic Explosions and Overthrusts," *Trans. Amer. Geophys. Union* (1933), pp. 238–42.

[2] A. C. Lawson, "Isostatic Compensation as a Cause of Thrusting," *Geol. Soc. Amer. Bull. 33* (1922), pp. 337–52.

"overthrust" refers to a large low-angle thrust in which the upper block has a greater displacement with respect to the center of gravity of the earth than has the lower block, the displacement of both blocks being in the same direction. Similarly the term "underthrust" is used to refer to a thrust in which the lower block has the greater displacement with respect to the earth's center of gravity, both displacements again being in the same direction. A thrust fault in which the displacement of the two blocks is in opposite directions would be a combination of the overthrust and the underthrust and thus does not require a separate consideration.

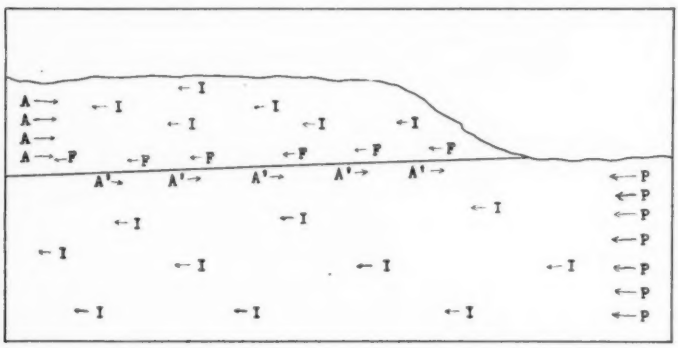

FIG. 1.—Forces involved in the overthrust. (For explanation, see below)

The forces involved in the mechanics of thrusting are here classified as follows: an *active* force is one which tends to move a portion of the earth with respect to the earth's center of gravity; a *passive* force is one which tends to resist such a movement.

Figure 1 represents the postulated mechanics of overthrusting. An active force is applied at the back of the upper block, tending to cause its translation to the right. This force is opposed by the friction on the thrust plane and by the inertia of the upper block, which are thus passive forces. When movement takes place, a portion of the active force is transmitted through friction to the lower block, where it is opposed by the inertia of the lower block and by its resistance to deformation. In order to have movement of the upper

block, it is necessary that it be strong enough to transmit a force great enough to overcome both the resistance on the thrust plane and the inertia of this block. Lawson[3] has calculated that, under what seem reasonable assumptions, the active force required at the back of the thrust plate to overcome the friction on the thrust plane alone will exceed the crushing strength of the strongest rock when the thrust plane reaches an extent of about 28 miles in the direction of movement. This estimate cannot be exact, but it serves to indicate an approximate limit to the extent of a thrust plane in the direction

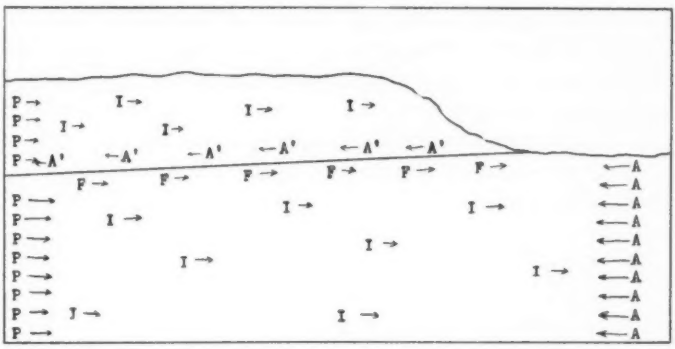

FIG. 2.—Forces involved in the underthrust

A, primary active force; *A'*, active force transmitted across thrust plane by friction; *I*, passive force of inertia; *F*, passive force of friction, and *P*, passive force of resistance of rock to deformation.

of slip. This conclusion was reached without considering the effect of the inertia of the upper block, which would tend to still further limit the extent of the thrust plane. This limit applies to the entire extent of the thrust plane, including both subaërial and subsurface portions. As the subaërial extent alone credited to some of the larger low-angle thrusts exceeds the limit set by Lawson, it is apparent either that the mechanics of an overthrust are not well understood or that these thrusts are not overthrusts.

Figure 2 illustrates the mechanics of underthrusting. In this case the active force is applied to one end of the lower block, tending to

[3] *Ibid.*, p. 341.

move it to the left. The major resistance to this force is the resistance of the other end of the lower block to deformation, which, owing to its differential value at the top and bottom of the block, sets up a shear, resulting in the fault fracture. After fracturing occurs, the mechanics change slightly. The friction on the fault plane acts as a minor resistance to the active force applied to the lower block, and through this friction an active force is transmitted to the base of the upper block. The magnitude of this force is determined by the frictional resistance on the thrust plane, and so is the same as the force resisting the movement of the upper block in the overthrust. In the underthrust the force transmitted across the fault plane is opposed by the resistance of the upper block to deformation and by its inertia. Without the inertia factor the limit set by Lawson would apply equally to the underthrust, as the upper plate would still be required to resist a force whose magnitude is determined by the friction on the thrust plane. The crushing strength of the rock in the upper plate would be exceeded with the same extent of thrust plane in the underthrust as in the overthrust.

The inertia of the upper block, however, is an additional passive force resisting the active force applied across the thrust plane. Thus the total resisting force is the sum of the strength of the upper block and of its inertia. If under certain conditions the value of the inertia factor is high, then that of the rigidity factor can be correspondingly low. Under any circumstances, however, the inertia of the upper block will tend to reduce the strength required in the upper block and so extend the limit of thrusting. In this way the inertia of the upper block favors the mechanics of underthrusting and opposes those of overthrusting.

If the inertia of the upper block is to be a factor in the mechanics of thrusting, it is necessary that the upper block acquire a motion relative to the center of gravity of the earth. In the overthrust this has always been assumed; but in the underthrust it has been generally considered that the upper block remains stationary, although nowhere has the writer been able to find mention of a point to which the movement on the thrust has been referred. If in the underthrust the upper block does move, then the relative movement between the blocks is only a portion of the actual crustal shortening.

It being established that under certain hypothetical conditions the inertia of the upper block will tend to favor the formation of large low-angle underthrusts, the writer has assumed the actuality of these conditions and has tried to determine the magnitude of this tendency.

An ideal case was assumed in which the upper block is like a brick resting upon a table. It is clear that, if the table moves slowly, the inertia of the brick will not be great enough to overcome the resistance of friction, and so it will remain stationary with respect to the table. If, however, the table moves with a high initial acceleration, the inertia of the brick will cause it to move less than the table, thus effecting a relative movement between them. The inertia of the brick has thus acted as a passive force tending to prevent the movement of the brick with respect to the earth's center of gravity. The extent of this relative movement is determined by three factors, the initial acceleration of the table, the rate of change of the acceleration of the table, and the friction between the brick and the table.

In the solution of this problem it was assumed that the rate of change of the acceleration of the table (lower thrust block) is a straight-line function of the time. It seems more probable that this change would be a parabolic function of the time; but as this would give a greater relative movement than is given by the straight-line change, the writer prefers to make the more conservative assumption.

The other factors of the problem are taken as varying between limits in order to cover all probable cases. The value of the coefficient of friction on the thrust plane is taken between the limits 0.0 and 1.0[4], and the initial acceleration of the lower block between the limits 8 and 512 feet per second per second. So far as the writer has been able to determine, nothing is known about the acceleration, the velocity, or the time involved in a single earthquake slip; consequently, the limits set are those which to him seem reasonable. If the reader considers that these limits are too restricted, he may extend the graph to fit his own conceptions.

[4] Lawson, *ibid.*, pp. 339–41; C. K. Wentworth, "The Russell Fork Fault of Southwest Virginia," *Jour. Geol.*, Vol. XXIX (1921), pp. 351–69.

In the graph (Fig. 3) the coefficient of friction between the upper and lower plates is plotted as the abscissa. The ordinate is the relative movement between the two blocks shown as a percentage of the

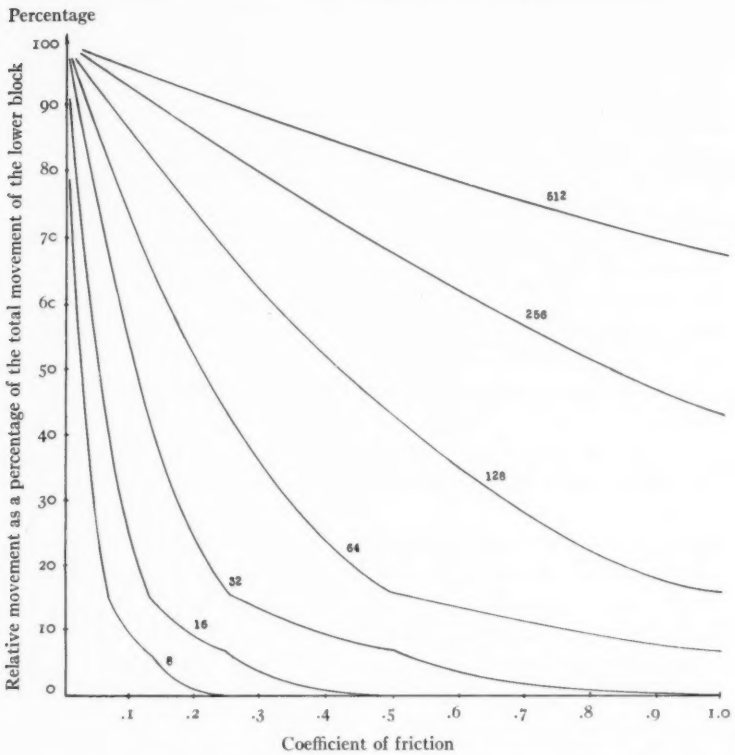

FIG. 3.— Relative movement as determined by the coefficient of friction and the initial acceleration The relative movement is plotted as a percentage of the total movement of the lower block. The figures on the graph lines show the acceleration in feet per second per second for that line.

total movement of the lower block. Separate curves have been drawn representing the percentage of differential movement as determined by the coefficient of friction and the initial acceleration of the lower block.

It is seen that in the case where the initial acceleration of the lower block is assumed to be 32 feet per second per second, the inertia of the upper block will not effect a relative movement between the two blocks if the coefficient of friction is greater than 1.0, and that the relative movement is not important if the friction is greater than 0.5. As the coefficient of friction is lowered, the relative movement increases greatly; and when the coefficient of friction reaches a value of 0.1, the relative movement is 50 per cent of the total movement of the lower block. If the initial acceleration of the lower block be increased, the differential movement between the two blocks for a given coefficient of friction will be a greater proportion of the total movement of the lower block. With an initial acceleration of 256 feet per second per second the differential movement between the two blocks is 42 per cent of the total movement of the lower block if the coefficient of friction is 1.0, 67 per cent if the friction is 0.5, and 93 per cent if the value of friction is 0.1.

It is thus seen that in an underthrust the inertia of the upper block may, under certain conditions, be a factor of great importance in causing relative movement between the two thrust plates. The presence of even a small additional passive force at the back of the upper plate would tend to greatly increase this relative movement.

For effective operation of inertia in the way described, the accumulated stresses must be relieved suddenly, thus giving the lower block a high initial acceleration. Rock fracture at the back of the lower block could cause this sudden release of stress.

As has been shown, the inertia of the upper block is a force which under favorable conditions can cause a relative movement between the two blocks. If a loose block on the surface of the earth has sufficient size so that the inertia will overcome the initial static friction (shear), it would be possible for it to undergo translation with respect to the material upon which it was resting. This translation would be greatly aided by a slope of the surface in the direction of the relative movement of the upper block. With the addition of the factor of initial slope of the surface over which the block is to move, these mechanics resemble those postulated by Reeves[5] for the Bear-

[5] Frank Reeves, "Thrust Faulting and Oil Possibilities in the Plains Adjacent to the Highwood Mountains, Montana," *U.S. Geol. Surv. Bull. 806* (1929), pp. 155–90.

paw and the Highwood Mountains. The difference lies in the directional character of the underlying movement and the major importance of the inertia of the upper plate.

As stated before, it has been necessary to make this analysis on the basis of assumed values for several factors because the true values are unknown. If more information could be obtained relative to the rate of change of acceleration, the duration of a single slip, and the maximum velocity attained in a single slip, the importance of inertia as a factor in the mechanics of thrust faulting could be more accurately determined.

The only conclusion justified at the present time is that if the mechanics of thrust faulting are such that inertia is a factor of importance, as it seems to be, then this factor will favor the mechanics of underthrusting and oppose those of overthrusting.

Acknowledgments.—The writer wishes to express his appreciation to Professors R. T. Chamberlin and F. M. Van Tuyl for their suggestions in the preparation of the manuscript and to Professor W. H. Bucher for discussions in the field that led to the clarification of the problem.

STRUCTURES IN THE DACITIC FLOWS AT CRATER LAKE, OREGON

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ABSTRACT

Flow structures and other features of the later lavas of Mount Mazama (Crater Lake), show that they had a viscosity between that exhibited by plug domes and that of "normal" lava flows. Three distinct flows are described, and evidence of the high gas content of the lava is given. The mechanics of their viscous outflow, as inferred from the structures, offers an explanation of the features of Rugged Crest. The "back-flow," heretofore cited as proof of subsidence of the mountain, is interpreted as the vent-filling for one of the flows.

INTRODUCTION

Among the latest lavas to be extruded from Mount Mazama, the volcano ancestral to Crater Lake, Oregon, were at least five separate flows which J. S. Diller, who mapped the area in 1886,[1] called dacitic in nature. Three of these lie on the north rim of the present lake, and a study of their structures during the summer of 1935 has led to conclusions as to their mode of origin. Two others, lying on the east rim of the lake, are yet to be studied.

The boundaries of these flows outside the rim are in most places masked by a thick pumiceous mantle, and are thus in part located physiographically. Two of the flows, however, are exposed in nearly complete cross section in the walls of the crater.

Observations were made of the following features in the Grouse Hill, Llao, and Cleetwood flows, wherever they were sufficiently exposed:

Flow structures:
- Banding (color, composition, texture)
- Parting along flow planes
- Chilled surfaces (glassy and ropy)
- Elongated vesicles and gas pockets
- Oriented phenocrysts

[1] J. S. Diller, *U.S. Geol. Surv. Prof. Paper 3* (1902).

Viscosity indications:

Configuration of the surface and base of the flows, and their included angles
Thickness, width, and length of the flows, and the ratios among them
Failure of lava to enter crevices
Degree of engulfment of underlying pebbles and boulders

Gas activity:

Quantity and size of included gas pockets, vesicles, etc., and their features
Evidences of refusion
Evidences of sublimation

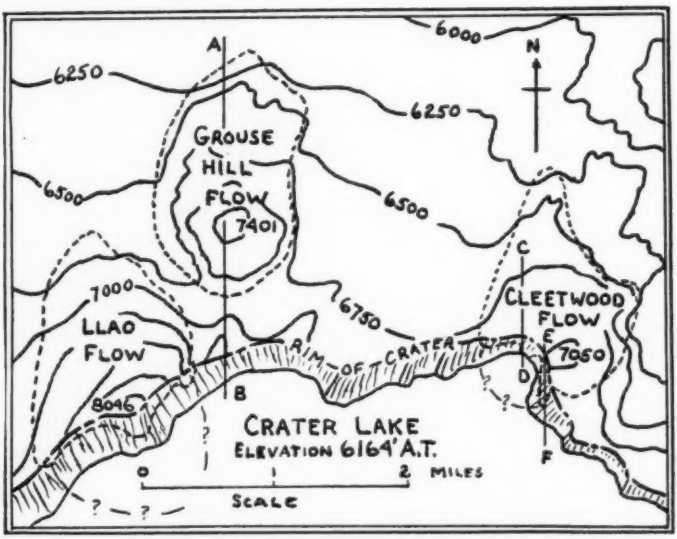

FIG. 1.—North rim of Crater Lake, showing dacite flows and location of cross sections.

GROUSE HILL FLOW

The Grouse Hill flow lies on a fairly smooth surface which slopes northward about 250 feet to the mile. It is almost 2 miles long, 1 mile wide, and averages 250 feet in thickness, rising, however, to 800 feet in the center. This point of highest elevation, probably indicating the location of the vent, is half a mile from the south or upper edge on the flow. The ratio of maximum thickness to breadth is here

about 1:7; and of thickness to length, 1:13. This would give a maximum calculated angle of about 16° between the base and surface of the flow.

This flow has suffered only a minor amount of erosion or other destruction; but although the more or less evenly sloping top is covered with a thick pumiceous layer, its structure is exposed in steep-walled cliffs on several sides. As one climbs the south wall, the flow planes, nearly horizontal at their base, are observed to dip inward (northward) more and more steeply, until toward the top they

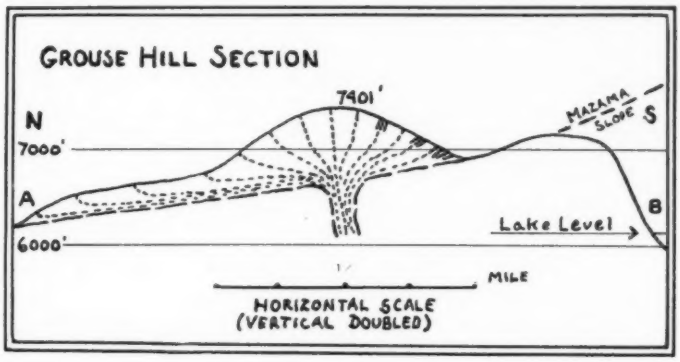

FIG. 2.—Cross section of Grouse Hill flow

approach verticality. Strikes of the flow planes are generally more or less parallel to the sides of the flow.

The north-south cross section *AB* of Figure 2 indicates actual and inferred (dotted) structures.

LLAO FLOW

Llao flow, which filled a glacial valley cut in the northwest side of Mount Mazama, and overflowed this valley on both sides, is 1,200 feet thick at the center (as exposed in the crater wall) and 400 feet thick at the south rim of the valley. It extends farther south from the rim of the valley for three-quarters of a mile, thinning as it goes.

Northward, the steep side of the filled glacial valley rises until the flow is only a hundred feet thick, but the lava spreads out laterally

for a quarter of a mile beyond this point before it pinches out. The width of the flow as exposed in the crater wall is thus a mile and a half, and its length from the present crater rim is probably only slightly more. Since the flow lies in a valley, an estimate of the thickness-breadth ratio lacks significance; but if both "wings" of the flow are included, the ratio is still only 1:6.

The flow structures are almost horizontal in the lower portion and steepen toward the upper surface, dipping toward the lake. In the upper 10–20 feet of the flow the structures are almost vertical, and the lava is distinctly dark and glassy.

In the newly exposed road cuts in the southwest flank of the Llao flow, the following features demonstrate a high viscosity of the lava, yet show that it was still liquid enough to move easily.

1. A boulder a foot in diameter protruding from the subjacent portion of the flow was overwhelmed, but the lava did not fill a 2-inch crevice at its sides.

2. Pebbles from 1 to 5 inches in diameter merely impressed themselves a third of the way into the base of the lava overflowing them.

3. Chilled, ropy (taffy-like) surfaces are very common on the basal contacts of the individual portions of the flow.

4. The highly banded lava in places was pulled apart along the flow planes, leaving glassy filaments stretched more or less completely across the cavities. Many of these filaments, instead of being sharply pointed, have globular tips, suggesting refusion.

5. Vesicles in some places make up as much as 30 per cent by volume of the rock. Most of them are elongated, but in few cases is their maximum diameter more than twice that of the minimum. Cavities with internal divisions are especially noticeable, and they may be up to a foot in diameter. A cave was discovered in Llao flow that measured 20 feet long and 10–15 feet in diameter, with many tributary channels a foot or so in diameter ramifying downward. The surfaces in the cave and channels varied from smooth glassy and ropy to extremely rough, sharp, and slaggy. Sublimed specular hematite crystals were found on the slaggy surfaces in one of the channels 30 feet below the surface.

CLEETWOOD FLOW

Cleetwood flow shows the same type of structure as does Llao in the cliffed walls on both sides of Cleetwood cove, with the important difference that the dips of the flow planes in the upper portions of both parts of the flow are inclined away from rather than toward the lake, as is the case at Llao. An estimate of the thickness of Cleetwood flow is not practicable; but the observed angle between the surface and base of the end of the flow exposed in the east

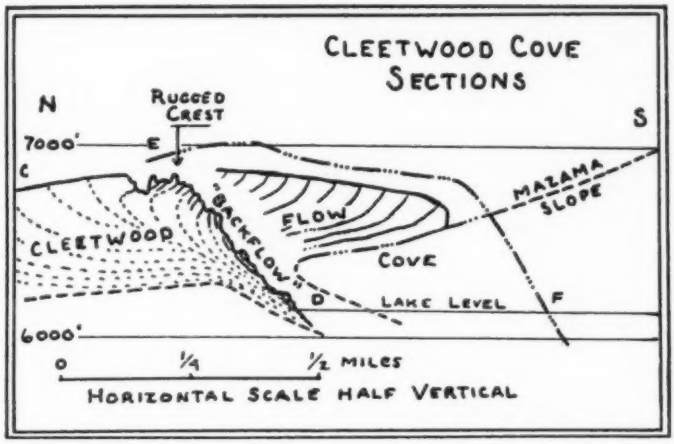

FIG. 3.—North-south cross sections of "backflow" (*left*) and east wall of Cleetwood cove (*right*).

wall of Cleetwood cove is 16°, which is the same as that calculated for Grouse Hill flow.

The lava of the "backflow" described by Diller,[2] the inferred structure of which is shown in Figure 3, has been very little disturbed except by normal erosion on a steep slope. The flow planes are distinct and quite continuous in attitude, even through the now more or less isolated blocks and pinnacles. The planes strike almost due east and west (which is true of nearly all the inclined flow planes in this

[2] *Ibid.*

area) and dip variously as shown in the figure, in which the "backflow" itself (left half of the diagram) appears in projection against the east cliff (right half), which really lies a quarter of a mile to the east. The west cliff (not shown) is similar.

Rugged Crest is an area just north of the "backflow," characterized by an east-west trending glen, from 50 to 300 feet wide, up to 100 feet deep, and perhaps half a mile long. It is studded with large, more or less isolated pinnacles of lava, usually bounded by smooth joint planes, some of the vertical or overhanging walls of which are at least 100 feet in height. The flow structures in these monoliths and in the cliffed walls of the glen, as well as in the pinnacles at the head of the "backflow," dip very gently to the north at the base and steepen to verticality toward the summit (Fig. 3). In the highest pinnacles, the lava at the top is glassy, indicating the remnants of the original surface of the flow. The well-known "Mazama Rock" on the rim highway illustrates all these features.

CONCLUSIONS

The structures in these three flows were to a great extent determined by the high viscosity of the lava, which prevented complete mixing of the various portions of the flow to form a homogeneous liquid and which promoted highly differentiated bands of various sorts. The flows were also shorter and steeper than usual; and the cooling at the surface must have been accelerated, with consequent formation of a chilled glassy porphyritic border of considerable rigidity. There was also a high gas content, in at least some parts of the flows.

The viscosity of the lava, which was sufficient to form in each case a steep-sided extruded mass with a thickness large in comparison to length and breadth, was not great enough to cause the lava to break up during extrusion into a plug dome of angular blocks and fragments, nor was it low enough for the lava to spread outward by the more familiar rolling and overturning motion of its front.

In the formation of one of these flows, the first highly viscous lava emitted from the center of extrusion was forced outward in all directions by newer lava welling up through its center and partially

overriding and partially displacing it as the later lava flowed outward. The weight of the overriding mass squeezed out that below to a thin layer.

This method of growth, although probably almost continuous, is shown simplified to four stages in Figure 4. Here it can be seen that the contacts of the lavas representing the different stages of growth are identical with the flow lines, or zones of shear, appearing in previous diagrams.

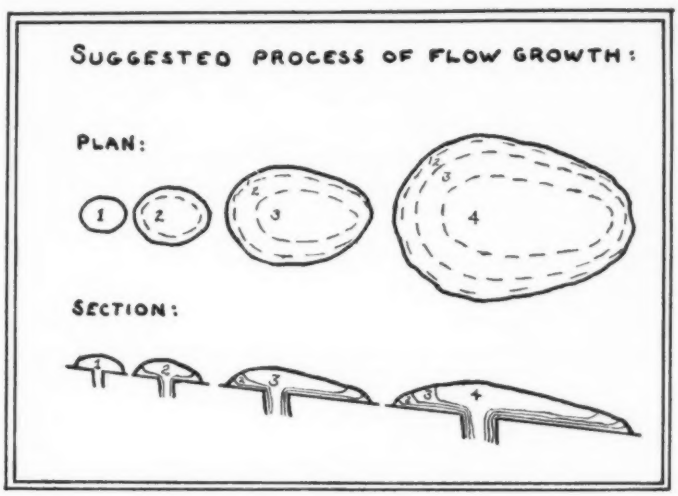

FIG. 4.—Four stages in the growth of a viscous flow

The quick cooling of the upper surfaces of the lava, soon after it was exposed to the air, froze the flow structures near those surfaces in positions more or less at right angles to them. Great sheets or blocks of this solid crust, with vertical structures near the upper surfaces, must often have been forced outward almost intact, riding upon the liquid beneath. Other portions of the crust were broken and isolated, tilted, or even overturned.

Rugged Crest glen seems to have been the result of the pulling-apart of a portion of this more rigid crust, producing a narrow and

deep defile with matched walls. Diller's[3] suggestion of a collapsed lava tube is unsatisfactory for a number of reasons.

Diller believed that Cleetwood flow was in the process of extrusion at the time of the collapse of Mount Mazama. The still motile lava poured back down into the newly formed cauldera, producing the structure he aptly labeled "backflow." This feature was used as evidence for the theory of collapse of Mount Mazama, as opposed to explosion. Figure 3 indicates that the "backflow" is simply the filling of the inclined vent of Cleetwood flow, which apparently came up along an east-west fissure dipping steeply southward into the prehistoric mountain and which has only subsequently become exposed in the crater wall. The position of the "backflow," midway between two projecting cliffs of similar lava with north-dipping flow planes, and the continuity of its structures from the rim of the present crater to the water's edge are all corroborative evidence in favor of this interpretation.

[3] *Ibid.*

A NEW SPECIES OF THE GENUS *PRISCACARA* FROM THE EOCENE OF WASHINGTON

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Some time ago Professor Joseph Daniels, of the University of Washington, forwarded to the University of California a specimen of a fossil fish that had been found about 1928 in a coal mine in central Washington. Since fossil fishes are rare on the West Coast, outside of the Miocene deposits of southern California, this fish was considered of special interest. Dr. Charles L. Camp, of the University of California, identified the specimen and made a series of notes and measurements on it. Unfortunately, other duties prevented him from carrying his work to completion and publishing his valuable observations. I wish to express my gratitude to Dr. Camp for placing his material at my disposal, and to Professor Joseph Daniels for his kindness and interest in forwarding the specimen a second time.

DISCUSSION OF THE GENUS PRISCACARA

The genus *Priscacara* was described by E. D. Cope in 1877,[1] but it was not until 1884 that specimens of the genus were figured.[2] Cope described, in all, seven species, based upon the beautifully preserved material collected in the Middle-Eocene Green River shales of southwestern Wyoming. He suggested that the genus was allied to the Pomacentridae but differed from the living forms in the possession of vomerine teeth and eight branchiostegal rays.[3] The existing Pomacentridae are marine. In the years following Cope's original description the family relationship of the Green River form was regarded as

[1] E. D. Cope, "A Contribution to the Knowledge of the Ichthyological Fauna of the Green River Shales," *U.S. Geog. Surv. Terr. Bull. 3* (1877), p. 816.

[2] Cope, "The Vertebrata of the Tertiary Formations of the West," *Rept. U.S. Geog. Surv. Terr.*, Vol. III, Book I (1884), pp. 1–1001.

[3] J. D. Haseman, "The Relationship of the Genus *Priscacara*," *Amer. Mus. Nat. Hist. Bull. 31*, p. 97.

uncertain. *Priscacara* was referred to the Pomacentridae, the Labridae, and the Cichlidae. Most writers followed the latter reference, but it was not until 1912 that the matter was definitely settled. J. D. Haseman[4] carefully listed the characters for and against the foregoing family relationships and demonstrated clearly that *Priscacara* was a cichlid fish.

PRISCACARA CAMPI, N. SP.

Type.—A complete, well-preserved skeleton of a fossil fish, in the collections of the College of Mines, University of Washington, Seattle, Washington.

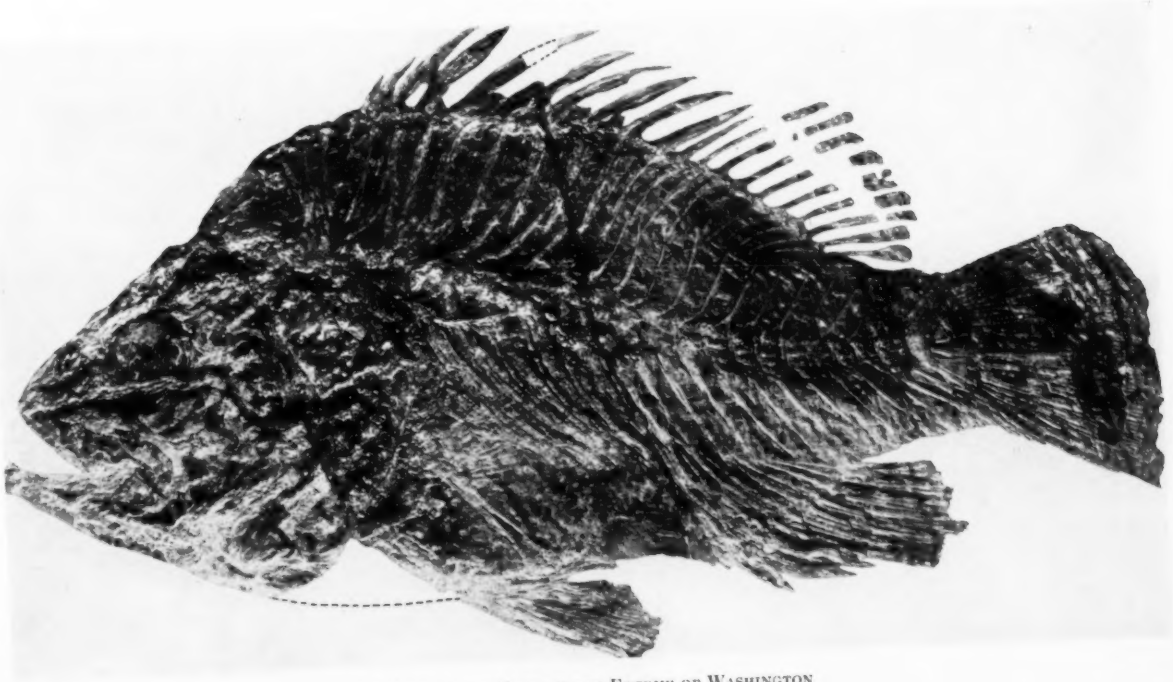
Type locality.—The specimen was donated to the College of Mines, University of Washington, by the Northwestern Improvement Company. It had been taken from the roof shales on the Eleventh West Entry in No. 2 Mine about 12 inches above the seam, Roslyn coal field, Kittitas County, central Washington (Mount Stuart quadrangle). The mine is in Section 5, T. 20 N., R. 14 E., and the workings extend south beneath Section 8 on the Roslyn coal bed. The elevation of the point from which the specimen came is 3,238 feet, in the upper workings of the Roslyn seam.

In general[5] the Roslyn-Clealum coal field comprises a series of Tertiary sediments deposited in an irregular basin, surrounded by and resting unconformably upon the pre-Tertiary complex. The age of these sediments was for some time in question, but considerable geologic investigation has enabled the age to be determined with accuracy. Tertiary sedimentation began with the Swauk formation, which unconformably overlies the older beds but was largely derived from them. Uplift and folding of this formation was followed by a period of volcanic activity, which resulted in the Teanaway basalt. Conformably overlying this basalt, and in part contemporaneous with it, is the Roslyn formation. The sediments of the Roslyn are alternating sandstones, shales, and clays, with several beds of coal, aggregating about 3,500 feet in thickness. The most productive coal beds are in the upper 1,200 feet of the formation, suggesting that

[4] *Ibid.*

[5] H. E. Saunders, "The Coal Fields of Kittitas County," *Wash. Geol. Surv. Bull. 9* (1914), pp. 51–60; G. O. Smith, "Mount Stuart Folio," *U.S. Geol. Surv. Folio 106* (1904), pp. 1–10.

PLATE I



Priscacara campi: FROM THE EOCENE OF WASHINGTON

deposition went on under steadily shallowing conditions. The specimen described here came from one of these coal beds in the Roslyn formation, in all probability the Roslyn seam itself. The Manastash formation, although not immediately overlying the Roslyn, is certainly later in time, but possibly represents the same cycle of deposition with but slight interruption. Fossil plants were collected in the mines of this region in the nineties, by members of the United States Geological Survey. Dr. F. H. Knowlton, after studying these floras, came to the following conclusions:[6] The Swauk was Lower Eocene, since it contained forms that were found in the Denver and Laramie formations on the Great Plains. The Roslyn was later, had no forms in common with the Swauk, had two in common with the Miocene of California, and the rest were new. The Manastash flora was of Clarno, or Upper Eocene, age.

The fossil fish here described as *Priscacara campi* accords in age with that set by Knowlton. The genus *Priscacara* is known only from the Green River shales, and the Washington species is not sufficiently different to suggest any but a Middle Eocene age.

Description.—The specimen is preserved in a fairly fine-grained, dark-gray, calcareous shale. This matrix has a very slaty appearance but may be easily scratched with the fingernail. Parts of the fourth and fifth dorsal spines are missing, and the lower part of the opercular region is broken away. It is a large fish (length, 255 mm.), deep bodied (114 mm., exclusive of the dorsal fin), with a long emarginate dorsal fin. The dorsal hump rises more sharply than in *Priscacara serrata*, resembling in general outline Cope's figure of *P. clivosa*[7] or Eastman's *P. dartonae*.[8] It falls very definitely into the first section of the genus noted by Cope,[9] on the basis of its large size and robust first ventral spine. Both dorsal and ventral spines are heavy, much as in *P. cypha*, but the development of the soft rayed fins is much greater than in the latter. The caudal peduncle is broader than in

[6] Smith, *op. cit.*, pp. 6–7; G. M. Wilmarth, *Tentative Correlation of the Named Geologic Units of Washington* (Washington: U.S. Geol. Surv., 1934), Chart I.

[7] Cope, "The Vertebrata of the Tertiary Formations of the West," *loc. cit.*, Pl. 13, Fig. 3.

[8] C. R. Eastman, "Fossil Fishes in the Collection of the United States National Museum," *Proc. U.S. Natl. Mus.*, Vol. LII (1917), Pl. 23.

[9] Cope, "The Vertebrata of the Tertiary Formations of the West," *loc. cit.*, p. 92.

P. serrata and extends beyond the margins of the soft dorsal and anal fins. This gives *P. campi* a more "stumpy" appearance. The caudal fin is not emarginate, but is less elongate than that of *P. dartonae* or *P. cypha.*

In contrast to *Priscacara dartonae, P. campi* is slightly smaller, has more robust spines, shorter caudal fin, soft rayed fins, less developed, and has the dorsal vertebrae less sharply flexed. *P. serrata* is smaller than *P. campi,* more elliptical in outline, has the anal fin inserted farther forward, and has a proportionately larger caudal fin but a smaller peduncle. The Washington form seems most closely allied to the foregoing Green River species.

Of the remaining Green River forms all except *Priscacara testudinaria* are smaller than *P. campi. P. testudinaria* is an elongate slender form, larger than any other member of the genus, and the only one described by Cope that was not from the Green River shales.

In summary, the characteristics of *Priscacara campi* are as follows: Radii: D X-11, AIII-8, C.18?, PII-?; vertebrae: D 8, C.13. First dorsal spine, one-half length of second; second, one-half length of third; third, very stout. Spines of anal fin heavy; first, one-half length of second. Other dimensions are given in Table I.

TABLE I

	Millimeters
Total length (tip of premaxillary to caudal fin)	255
Greatest depth (at third dorsal spine)	114
Axial length of skull	85
Length to base of first dorsal spine (from Pmx.)	110
Length to base of first dorsal spine soft ray	115
Length to base of caudal fin	204
Depth at orbit	64
Depth of caudal peduncle	36.5

Orbit, five in length of skull; skull, three in length of body.

DISTRIBUTION OF THE CICHLIDAE

The family Cichlidae is well known in the modern fish faunas. Its forms are considered fresh water, although it is not impossible that they inhabit brackish water and may maintain themselves for some time in a marine environment.[10] Today the family is known

[10] G. S. Myers, "The Existence of Cichlid Fishes in Santo Domingo," *Copeia,* No. 167, p. 35; Haseman, *op. cit.,* p. 99.

from over eighty-eight genera and five hundred species. These forms are characteristic of South America (27 genera), Africa (58 genera), Madagascar (3 genera), and the Indo-Syrian region. The question of the origin of this peculiar Ethiopian-Neotropical distribution of the family Cichlidae is one that has long puzzled ichthyologists. It was thought necessary to derive this dispersal by means of a former land connection between South America and Africa, either through Antarctica or by some sort of Gondwana land.

The foregoing question makes the fossil record of this family important, for this record may offer the solution of the problem of the origin of this tropical distribution. The first fossil occurrence to be definitely placed in this family was that referred to the existing genus *Acara*, collected in the Tertiary lignite, in the province of Saõ Paulo, Brazil.[11] In 1907 Sauvage[12] reported the occurrence of a cichlid fish in the Miocene of Oran, Algeria. This was the first record of the family outside the area of its modern distribution. Sauvage gave these fossils a new generic name, *Palaeochromis;* but Woodward does not support this reference,[13] and regards the specimens as *Chromis*. Haseman, working on the distribution of South American fishes, felt that it was necessary to settle the reference of *Priscacara* definitely. In 1912 he proved conclusively that this North American genus was a member of the Cichlidae, although probably not directly ancestral to the modern fauna. *Cichlasoma woodringi,*[14] a fossil cichlid from the Miocene of Haiti, is stated by its author to be close to the living *C. tetracanthus*, an observation with which Myers agrees.[15]

From the foregoing meager and scattered fossil record one may deduce but little in the way of definite facts. It would seem that the North African and Haitian records would indicate a northward extension of the range of this family in the past. Then, too, these Mio-

[11] A. S. Woodward, "Consideracoes sobre alguno peixes terciarios dos schistos de Taubata, Estado de S. Paulo, Brazil," *Revista Mus. Paulista*, Vol. III (1898), p. 69.

[12] H. E. Sauvage, "Sur des poissons de la famille des Cichlidés trouvés dans le terrain tertiare de Guelma," *Comptes Rendus Acad. Sci. Paris*, Vol. CXLV (1907), p. 360.

[13] Karl A. von Zittel, *Textbook of Paleontology*. Revised by A. S. Woodward (New York: Macmillan & Co., 1932), Vol. II, p. 174.

[14] T. D. A. Cockerell, "A Fossil Cichlid Fish from the Republic of Haiti," *Proc. U.S. Natl. Mus.*, Vol. LXIII (1923), p. 2.

[15] Myers, *op. cit.*, p. 35.

cene records are most closely allied to modern genera, pointing to an Eocene or Oligocene origin of the modern forms. If this is correct, then *Priscacara* would represent a side branch of the family, since one would expect to find Middle Eocene forerunners of the family more like the modern genera. *Priscacara*, then, does represent an older phase of the Holarctic distribution of the Cichlidae. The form, *P. campi*, described in the pages above, is the first occurrence of this genus outside the Green River area. It suggests that this and probably allied genera were widespread and common in Eocene times, but lack of fresh-water Eocene deposits has made the solution of the foregoing problem difficult.

A *SPORADOCERAS* FROM AMERICA

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ABSTRACT

The Devonian ammonoid genus *Sporadoceras*, though widespread and abundant in the Eastern Hemisphere, has not been recognized previously in the Western Hemisphere. A typical representative of it from the Upper Devonian Conewango group of northwestern Pennsylvania is described and illustrated in this paper; the distribution of the genus is discussed; and it is concluded that the Conewango group is probably, in part at least, correlative with the *Prolobites-Platyclymenia*-Stufe (Oberdevonstufe III) of Germany.

In their recently issued monograph on the cephalopods of the Upper Devonian Conewango group of New York and Pennsylvania, Flower and Caster[1] described a single goniatite. They recognized its uniqueness and coined for it the specific name *milleri*, referring it with question to the late Paleozoic genus *Paralegoceras*. Further study, however, has shown that the affinities of this specimen are rather with the Upper Devonian genus *Sporadoceras*, which has never before been found in this hemisphere. It was through the courtesy of Professor A. C. Swinnerton, of Antioch College, that we were able to study this specimen; and we wish to take this opportunity to express our sincere appreciation for this favor.

The genus *Sporadoceras* was established in 1883 by Hyatt,[2] who designated as its genotype *Goniatites bidens* Sandberger and Sandberger of the Upper Devonian of Germany, which may be the same as *Ammonites* [*Sporadoceras*] *münsteri* von Buch of the same general horizon and locality. Frech, in his classical study *Über Devonische Ammoneen* and in his catalogue *Ammoneae Devonicae*, indicates that

[1] R. H. Flower and K. E. Caster, "The Stratigraphy and Paleontology of Northwestern Pennsylvania, Part II: Paleontology, Section A: The Cephalopod Fauna of the Conewango Series of the Upper Devonian in New York and Pennsylvania," *Bull. Amer. Paleon.*, Vol. XXII (1935), pp. 56–57, Pl. VIII, Figs. 1, 2.

[2] Alpheus Hyatt, "Genera of Fossil Cephalopods," *Boston Soc. Nat. Hist. Proc.*, Vol. XXII (1883), pp. 321–22.

congeneric forms have been found at many localities in Germany and in France, England, and Russia (southern Urals); and Haug[3] adds Sardinia to this list. Sobolew[4] has described and illustrated excellent representatives of *Sporadoceras* from Poland, and Schmidt[5] indicates that the genus is represented also in Spain (the Pyrenees). Von Peetz[6] states that in Siberia (Kirghiz Steppe) Meister and Krasnopolsky found "*Goniatites* (*Tornoceras*) cf. *bifer* Sow." = *Sporadoceras* cf. *S. biferum* (Phillips); Haug[7] and Menchikoff[8] indicate that, locally at least, *Sporadoceras* is abundant in Northern Africa; and Delépine[9] has recently described excellent representatives of it from Western Australia. Truly, then, the genus is widespread in the Eastern Hemisphere, occurring in Europe, Asia, Africa, and Australia; and it has always seemed strange that we did not find it in North America—no Devonian ammonoids of any kind are known from Central or South America.

The stratigraphic range of the genus in Europe, at least, is well known. There the Upper Devonian is divided into six ammonoid zones which the German paleontologists and stratigraphers (partic-

[3] Émile Haug, *Traité de géologie, Tome II: les périodes géologiques* (Paris, 1908), p. 703.

[4] D. Sobolew, "Skizzen zur Phylogenie der Goniatiten," *Mitt. Warschauer polytechn. Inst.* (1914), pp. 18, 24, 32–40, 45, Pl. I, Figs. 10, 11; Pl. II, Fig. 5; Pl. III, Figs. 2, 3, 17, 18, 21; Pl. V, Fig. 12; Pl. VI, Fig. 5.

[5] Hermann Schmidt, "Das Paläozoikum der Spanischen Pyrenäen," *Abhandl. Gesellsch. der Wissenschaften zu Göttingen, math.-physik. Klasse*, Folge III, Heft 5 (1931), p. 50.

[6] H. von Peetz, "Beiträge zur Kenntniss der Fauna aus den devonischen Schichten am Rande des Steinkohlenbassins von Kusnetzk," *Travaux Sect. Géol. Cab. Sa Majesté*, Vol. IV (1901), pp. 315, 386.

[7] Émile Haug, "Sur deux horizons à céphalopodes du Dévonien supérieur dans le Sahara oranais," *Comptes Rendus Acad. Sci. Paris*, Tome CXXXVII (1903), pp. 83–85; *Traité de géologie*, Tome II: *les périodes géologiques* (Paris, 1908), pp. 722–24.

[8] Nicolas Menchikoff, "Recherches géologiques et morphologiques dans le Nord du Sahara occidental," *Revue géog. phys. et géol. dyn.*, Vol. III (1930), pp. 140–44; "Sur le Dévonien à céphalopodes de l'Oued Saoura et des chaînes d'Ougarta (Sahara oranais)," *Comptes Rendus Acad. Sci. Paris*, Tome CXCIV (1932), pp. 1966–68; "La série primaire de la Saoura et des chaînes d'Ougarta," *Gouvt. Gén. l'Algérie, Bull. Service la Carte Géol. l'Algérie, 2e sér., stratigraphie, travaux récents des collaborateurs*, Fasc. II (1933), pp. 116–17.

[9] Gaston Delépine, "Upper Devonian Goniatites from Mount Pierre, Kimberley District, Western Australia," *Quart. Jour. Geol. Soc. London*, Vol. XCI, Part II (1935), pp. 208–15, Pls. XII, XIII.

ularly Wedekind and Schindewolf) term the "*Manticoceras-*, *Cheiloceras-*, *Prolobites-Platyclymenia-*, *Laevigites-Gonioclymenia-*,[10] *Wocklumeria-*, und *Gattendorfia*-Stufe," or merely "Oberdevonstufe I [=*Manticoceras*-Stufe]," "Oberdevonstufe II [=*Cheiloceras*-Stufe]," etc. According to Wedekind[11] and Schindewolf,[12] the genus *Sporadoceras* makes its appearance in the upper half of the zone of *Cheiloceras*, becomes particularly abundant in the next zone, the *Prolobites-Platyclymenia* zone, dwindles in the overlying *Laevigites-Gonioclymenia* zone, but extends up into the zone of *Gattendorfia*. All of the forms that occur in the *Cheiloceras* zone belong in the group of *Sporadoceras biferum* (Phillips), which is characterized by the fact that in the mature sutures the lobes next to the ventral lobe (the second adventitious lobes) are rounded (rather than pointed) and in most cases are distinctly shorter than the adjacent lobes on the lateral zones of the conch (the first adventitious lobes). Most of the forms that occur in the *Prolobites-Platyclymenia* and overlying zones belong in the group of *Sporadoceras münsteri* (von Buch) in which both pairs of lobes on the lateral (including ventro-lateral) zones of the conch are pointed and the lobes next to the ventral lobe are as long or longer than the adjacent lobes on the lateral zones of the conch. A few representatives of the group of *S. biferum* are known from the *Prolobites-Platyclymenia* and the *Laevigites-Gonioclymenia* zones, but no representatives of the group of *S. münsteri* have been found in the zone of *Cheiloceras*. Within the group of *S. biferum* we may recognize with Wedekind a subgroup of *S. sedgwicki* Wedekind in which the two pairs of lobes on the lateral (including ventro-lateral) zones of the conch are of about the same length—this subgroup includes, besides its type species, only *S. tenuidiscus* Schindewolf and the American form under consideration, *S. milleri* (Flower and Caster). The only known representative of *S. sedgwicki* came

[10] The generic name *Laevigites* is probably to be suppressed as a synonym of *Clymenia*, s.s., which has priority.

[11] Rudolf Wedekind, "Die Genera der Palaeoammonoidea (Goniatiten), (mit Ausschluss der Mimoceratidae, Glyphioceratidae und Prolecanitidae)," *Palaeontographica*, Band LXII (1918), pp. 147–49.

[12] O. H. Schindewolf, "Beiträge zur Kenntnis des Paläozoicums in Oberfranken, Ostthüringen und dem Sächsischen Vogtlande," *Neues Jahrb. f. Min., Geol. u. Pal.*, Beilage-Band XLIX (1923), pp. 339–52.

from the upper half of the zone of *Cheiloceras* (Oberdevonstufe IIβ) at Enkeberg, Germany, whereas all of the known representatives of *S. tenuidiscus* came from the upper half of the zone of *Prolobites-Platyclymenia* (Oberdevonstufe IIIβ) at Gattendorf, Germany. *S. milleri* is perhaps a little closer to *S. tenuidiscus* than to *S. sedgwicki*, and this might be taken to indicate that it came from beds which are correlative with Oberdevonstufe III rather than Oberdevonstufe II. Only one other goniatite has been found in eastern North America above the zone of *Manticoceras* (Oberdevonstufe I)—it is the holotype of *Tornoceras* (*Tornoceras*) *edwin-halli* Clarke, and it came from the Conneaut group which immediately underlies the Conewango group and immediately overlies beds that contain *Manticoceras* (Canadaway group). It therefore seems quite possible that the Conneaut is the approximate stratigraphic equivalent of the German Oberdevonstufe II and that the Conewango (or possibly only the lower portion of it) corresponds to Oberdevonstufe III—the available data in regard to the goniatites, at least, are compatible with these tentative correlations, but they cannot be said to establish them. It should perhaps also be mentioned in this connection that the goniatite fauna of the Three Forks shale of Montana has recently been restudied by Schindewolf,[13] who correlated the Three Forks formation with Oberdevonstufe III.

It was formerly believed by some paleontologists and stratigraphers that the Conewango might be Mississippian, or at least Devono-Carboniferous, in age; but the goniatite we are studying is clearly a Devonian type, and not a particularly late one at that. It is probably either of about the same age as or very slightly older than the goniatites of the Three Forks shale, and in Germany three younger Devonian ammonoid zones are recognized above the equivalent of the Three Forks.

DESCRIPTION OF SPECIES

SPORADOCERAS MILLERI (FLOWER AND CASTER)

Holotype, only known representative of this species, is a moderately well-preserved internal mold representing right half of adoral

[13] O. H. Schindewolf, "Über eine oberdevonische Ammoneen-Fauna aus den Rocky Mountains," *Neues Jahrb. f. Min., Geol. u. Pal.*, Beilage-Band LXXII, Abt. B (1934), pp. 331–50.

five camerae of phragmacone and adapical portion of living chamber of a mature individual. Conch moderately large, ammoniticonic in mode of growth, and sublenticular in shape. Whorls compressed, flattened (but distinctly convex) laterally, rather narrowly rounded ventrally, and deeply impressed dorsally. Lateral sides of conch round evenly into ventral side. At maturity whorls appear to be almost twice as high as wide; but as holotype is incomplete laterally and is slightly crushed, width of whorls cannot be determined with precision. At adoral end of phragmacone of holotype conch is about 50 mm. high. Depth of impressed zone is equal to about one-half height of conch. Umbilicus small, closed, and inconspicuous. Umbilical shoulders low, rounded, and indefinite. Traces of growth-lines are clearly discernible on holotype; these are of the so-called convex type, as each forms a moderately deep, rounded ventral or hyponomic sinus and a single low, broad, broadly rounded salient (extending from ventro-lateral zone of conch to umbilicus) on each of the broad flattened lateral zones of conch.

Camerae moderate in length (Fig. 1, *1*)—adoral camera of holotype is distinctly shorter than preceding camerae, indicating that specimen represents a mature individual. As shown by Figure 1, *3*, at maturity each suture forms a rather large subangular V-shaped ventral lobe, and on either side of it a rounded U-shaped ventro-lateral saddle, a similar but slightly asymmetrical ventro-lateral lobe (a second adventitious lobe), a narrower higher rounded tongue-shaped saddle, a larger deeper asymmetrical curved pointed V-shaped lobe (a first adventitious lobe), a broad relatively broadly rounded dorso-lateral saddle, a smaller narrowly rounded more or less V-shaped lobe (a first lateral lobe) on the umbilical seam, a rather broad slightly asymmetrical rounded U-shaped internal second lateral saddle, a narrower deeper pointed tongue-shaped internal lateral lobe, and a similar but rounded internal first lateral saddle which extends to the dorsal lobe—the shape of the dorsal lobe cannot be ascertained from the holotype, but presumably it does not differ greatly from that of other congeneric forms. Siphuncle small, circular in cross section, and ventral and marginal in position.

Observations.—In so far as form of conch, nature of umbilicus, and shape of sutures are concerned, the genera *Posttornoceras* Wedekind

and *Sporadoceras* Hyatt, which belong in different superfamilies, do not differ essentially; but the growth-lines of *Posttornoceras* are of the biconvex type, that is, they form lateral (as well as ventral) sinuses, whereas those of *Sporadoceras* are of the convex type, forming lateral

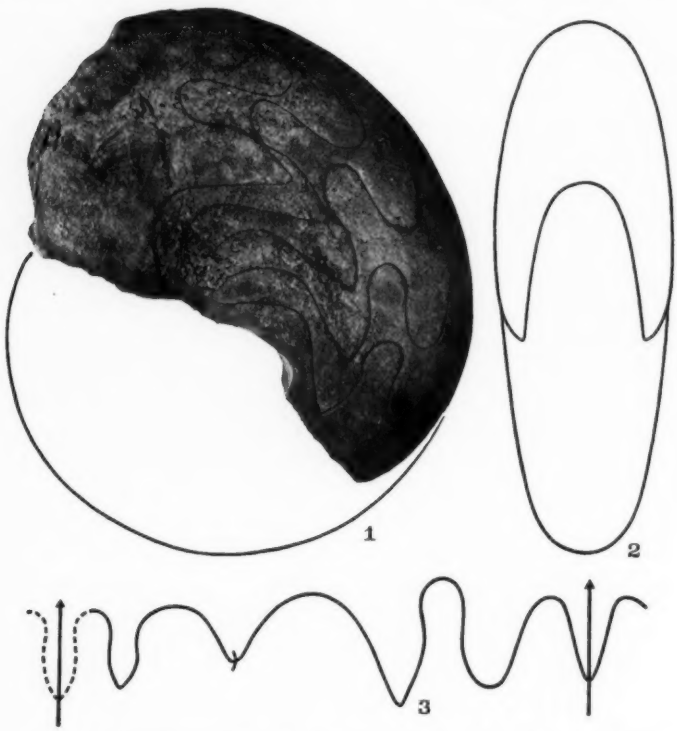

FIG. 1.—The holotype of *Sporadoceras milleri* (Flower and Caster), ×1. Illustration *1* is a retouched photograph; Illustration *2* is a restoration; and Illustration *3* is a diagrammatic representation of one of the adoral sutures.

salients rather than sinuses—it would be difficult to point out a better example of convergent evolution than is presented by these two genera. Traces of the growth-lines on the holotype of the species described above indicate clearly that its affinities are with *Sporadoceras* and not *Posttornoceras*. It is perhaps more closely related to *S.*

tenuidiscus Schindewolf and *S. sedgwicki* Wedekind, both of the Upper Devonian of Germany, than to any other described species. In the sutures of both of these species, the lobes next to the ventral lobe are much wider than in the form under consideration. Also, in *S. tenuidiscus* the lateral sides of the conch are nearly flat, they converge ventrad, the ventral zone of the conch is very narrow, and the whorls are subsagittate in cross section. *S. inflexum* Wedekind, also of the Upper Devonian of Germany (Oberdevonstufe IIIβ) is somewhat similar to *S. milleri*, but in its sutures the lobes next to ventral lobe are very narrow and are curved toward the venter.

Occurrence.—Probably lower Conewango, near the horizon of the Panama conglomerate, Howard Quarry, Erie County, Pennsylvania.

Holotype.—E. J. Armstrong Collection, Antioch College, Yellow Springs, Ohio.

REVIEWS

Geologie von Grönland. By LAUGE KOCH. Berlin: Gebrüder Borntraeger, 1935. Pp. 159; figs. 12.

A general work on the geology of Greenland is much needed and will be widely welcomed, in view of the epoch-making studies since Böggild's "Grönland" (in *Handbuch der regionalen Geologie* [1917]). When Böggild's book came out it was necessary to show on the geological map north of the area of basalt in Middle Greenland two formations only—Grundgebirge and undifferentiated sediments. With slightly more of detail this was true also of the Danish map of 1921 (*Grönland, i tohundredaaret for Hans Egede's Landing:* Atlas [Copenhagen: Reitzel, 1921]).

That Koch is now able to print reconnaissance geological maps for the entire northern half of Greenland showing definite horizons throughout, and all fixed in age by fossil contents, is almost entirely due to the work of Koch himself, and, within the last few years for East Greenland, with the geologists of his own expeditions. To have accomplished this great task almost unaided and under conditions of pioneer exploring within one of the two regions of the globe most difficult of access has no parallel in the long history of geology. Too great praise can hardly be accorded it.

For each geological map prepared it has been necessary first to provide an areal map based on key points fixed by astronomical observations and a triangulation, for preparing which Koch developed a technique specially adapted to the conditions. Already in the years 1917–23 he was able to prepare for the printer an areal map of all North Greenland on the scale of 1:300,000 (*Map of North Greenland* [Geodetic Institute of Denmark, 1932]), and the work has now been extended to the area of East Greenland by Koch's latest expedition.

A "Geology of Greenland" should beyond question have included a summary of the geology long ago worked out for South Greenland, but this it fails to do. On this earlier work the well-known geologists Böggild, Steenstrup, Nordenskjöld, and Ussing were engaged, and the igneous rocks particularly have great interest and have been particularly well studied. This serious omission, together with an alleged looseness of statement as well as definite misstatements of fact, with a claim that credit has not been given where due, etc., has been pointed out in a state-

ment signed by eleven members of Koch's large corps of geologists in the latest expedition ("Remarks upon Lauge Koch; *Geologie von Grönland*, 1935," *Med. fra Dansk Geologisk For.*, Vol. VIII [1935], pp. 497–511). Though some of their statements seem to be verified as regards details, others are trivial, and they are made in a particularly bad spirit, marked by an entire failure to recognize the importance of the larger results of their leader's work, and this is especially true of their list of Koch's published works on Greenland. Between the lines one may almost read a condition of near-mutiny not wholly unknown in connection with polar expeditions although seldom so fully aired.

Some inaccuracy as regards detail is probably connected with a hurried summary of a vast subject, and in his Preface Koch explains:

> It is a nearly impossible task to make correct reference to the different works on the geology of East Greenland, since some of the authors have worked only on short summer journeys, while others have made more thorough investigations. Further, the East Greenland studies are in such a state of rapid development that when as leader of the Expedition and editor of publications I completed the manuscript on July 1st, 1934, in tracing the line of development I could not too much emphasize the authors who worked on the different problems.

To the great development of the pre-Cambrian in Greenland, Koch gives the name "Greenlandium," and he gives reason for thinking the Greenland shield is distinct from the Canadian one. In the great development of the hitherto largely undifferentiated sedimentary series in North and East Greenland, starting upon a pre-Cambrian tillite, Koch has made out the formations of the following stratigraphic succession fixed by the fossil remains: Cambrian, Ozarkian, Ordovician, Gotlandian, Devonian (Middle and Upper), Carboniferous (Dinantian and Namurian), Permian (Lower), Zechstein, Triassic, Rhaetic, Lias (Lower, Middle, and Upper), Jura (Dogger and Malm), Cretaceous (Lower and Upper), and Tertiary (poor in fossils but in three series). To have contributed to the science of geology such a stratigraphic succession in a vast unstudied landmass of the globe is a geological event of the first order of magnitude to mark our generation.

The geological history that has developed about the Greenland shield starts, according to Koch, with the stable period of the Greenlandium, followed by glaciation on the East, and in turn by the Scanian period of great eruptive activity along both the north and the east coasts throughout an extent of 2,000 kilometers, with extrusions largely acid toward the west and basic on the east.

There followed a transgression by the Cambrian sea and the subsequent stable period of Cambrian-Ordovician, to which succeeded folding movements which were added to the Caledonian movements which had folded the sediments of the north and east coasts in the Smith Sound and East Greenland geosynclines, which meet in a vertex northeast of Peary Land in the northeast corner of Greenland.

At the close of the Caledonian folding Greenland lay very high and underwent further movements throughout the Devonian and the greater part of the Carboniferous and Lower Permian. There followed in East Greenland the first great marine transgression since the Caledonian time, with smaller regressions and new transgressions. During the Tertiary, movements were resumed on a large scale, though with noteworthy differences for different parts of the east coast.

At the conclusion of his book Koch prints the geological literature, which includes 123 items.

WILLIAM H. HOBBS

Petroleum Investigation. Hearings before a Subcommittee of the Committee on Interstate and Foreign Commerce. House of Representatives, 73rd Congress, H. Res. 441, Part II. Washington: Govt. Printing Office, 1934. Pp. 522; figs. 131. $1.75.

This is but one of five parts (the last an index) of the printed results of this investigation, but few geologists will find much of interest in the other parts. Part II contains three papers, the first by the United States Geological Survey, the others by the United States Bureau of Mines, and is free from the numerous testimonies characteristic of the other parts. From the point of view of the public it would seem that a few dollars might have been saved if, instead of printing the first paper of Part II, each interested member of Congress had been presented with a complimentary copy of Emmons, or supplied with free tuition to a home-study course in the subject.

The first paper starts out with an excellent twenty-seven-page summary of the history of petroleum given in six sections by David White. This is followed by a brief section of fifteen pages on the origin of petroleum by the same writer. Section 8, by several members of the Survey staff, is on "Petroleum Geology Summarized by States"; it consists of 161 pages with 130 figures, many of them folded insets, four of them in colors. Most of these are from published sources; conspicuous among those made especially for this volume are the generalized geologic columns showing the

oil occurrences of the various states. On the whole this portion of the volume is recommended highly to non-specialistis, including students who wish to get a bird's-eye view of geologic conditions in the oil fields of the United States.

Figure 1 is a colored map of the United States showing impossible, improbable, unfavorable, and possible petroliferous areas. In spite of several errors in printing, it is of considerable interest, expecially in contrast with the corresponding map in Arnold and Kemnitzer, *Petroleum in the United States* (1931). Contrasting Figure 1 with Figure 2 (oil fields of the United States) one is struck by the lack of consistency in that what appears as part of the Gulf Coastal province of Figure 1 is put with the Mid-Continent province of Figure 2. This is presumably because the United States Geological Survey feels that the precedent of not dividing the Gulf Coastal fields into inner and outer regions is too great to be overcome, or at least that a publication such as that under review is not the one in which to advocate a change.

In section 9 of eleven pages H. B. Soyster discusses estimates of oil reserves, and concludes that the United States reserves as of December 31, 1933, were 13.36 billion barrels. The paper closes with a brief description of petroleum on public and Indian lands.

The second paper is on "Petroleum Production and Development," by H. C. Miller and B. E. Lindsly. After taking up exploration, leasing, well spacing, and drilling, the writers describe the removal of oil by means of wells, following which they discuss waste, unit operation, and proration. Although the treatment of so many subjects in 220 pages requires brevity, many references are listed in the 227 footnotes, and the subject is excellently handled considering its purpose.

The final paper of seventy-eight pages by A. J. Kraemer on "The Effect of Technologic Factors on Supply of and Demand for Petroleum Products," in addition to covering what would be expected from the title, contains a brief description of petroleum refining methods and a longer section on alternative fuels.

D. JEROME FISHER

"Outline of the Geology of the Musoma District," by G. M. STOCKLEY, in *Tanganyika Geological Survey Bulletin 7*. Dar Es Salaam: Government Printer, 1935. Pp. 64; pls. 3; maps 4. 4*s*.

This report deals with an area of over 2,000 square miles lying immediately east of Lake Victoria. The oldest rocks of the region comprise several series of pre-Cambrian metasedimentary and metaigneous rocks.

Among these are banded ironstones which are cherts with bands of limonite and hematite and are interpreted as felsitic pyroclastics that have been partly replaced by silica, carbonates, and iron minerals through the agency of solutions from granitic intrusives.

There follow unconformably younger sediments of the Ikorongo series, which may, though unfossiliferous, be Paleozoic rather than pre-Cambrian. Tertiary phonolites and Pleistocene and Recent superficial deposits complete the sequence.

Igneous intrusions, dominantly large masses of granite, outcrop in most parts of the region. Innumerable dike rocks penetrate the pre-Cambrian granites and metamorphic rocks but do not traverse the younger rocks. The granites show some age diversity but are all pre-Cambrian.

Gold mineralization appears to be genetically associated with the youngest of the granites. It takes the form of (1) quartz veins, which are regarded as pegmatitic offshoots from the granite and occur mainly in the country rocks but subordinately in the granite itself; (2) impregnations of the country rocks of the Musoma series by solutions from the granite; (3) mineralized dikes of aplite, etc.; (4) impregnations of the country rocks by quartz stringers; and (5) rubble deposits.

The primary minerals of the quartz veins are gold, pyrite, chalcopyrite, arsenopyrite, galena, epidote, and tourmaline. There has been residual enrichment in gold in the oxidized zone. The writer also believes that there has been some solution of gold in the oxidized zone through the agency of lateritic manganese oxides which are of common occurrence. Such dissolved gold was redeposited under reducing conditions below.

Impregnations of the country rock with auriferous pyrite occur near certain granitic apophyses. A few aplitic dikes, apparently offshoots from the granite, carry auriferous pyrite and arsenopyrite. Rubble deposits are residual lateritic deposits associated with the veins. Some of their gold may be the result of solution of vein gold in the presence of manganese oxides and some is purely residual.

E. S. B.

Die Lagerstätten der Edelsteine und Schmucksteine. By O. STUTZER and W. F. EPPLER. Vol. VI of "Die wichtigsten Lagerstätten der Nicht-Erze." Berlin: Gebrüder Borntraeger, 1935. Pp. xvii+567; figs. 154. Rm. 45.50, bound.

The first 216 pages by the senior author is a completely re-written second edition of his work on the diamond. The remainder, by the junior

author, covers precious and semi-precious stones arranged in chemical-mineralogical order as follows:

Class	Representatives	Pages
Oxides	Silica, zircon, corundum, hematite	108
Haloid	Fluorite	1
Carbonates	Malachite, azurite	3
Aluminates	Spinel, chrysoberyl	6
Phosphates	Turquois, variscite	19
Silicates	Tourmaline	17
	Topaz, euclase	12
	Olivine, scapolite	4
	Lapis lazuli	4
	Garnet, dioptase	19
	Spodumene, rhodonite, jade	26
	Beryl	58
	Feldspar	10
Titanate	Benitoite	2
Organic compounds	Amber, jet	31

Locality and subject indexes comprising 27 pages complete the volume.

The plan followed in general for each mineral is to give a summary of its mineralogical properties, followed by detailed descriptions with production figures of the various deposits of economic importance for each variety; then comes brief treatment of the economically unimportant deposits, with finally a short section listing prices and grades. The work differs from what may be called "mineralogical gem books" such as Bauer-Schlossmacher's *Edelsteinkunde* (3d ed., 1932) and Eppler's *Edelsteine und Schmucksteine* (2d ed., 1934) in that it is devoted primarily to descriptions of geologic occurrences; it is an economic geology of precious stones.

For the English reader, the works of Wagner and the two Williams' on the South African diamond occurrences are more valuable than the 108 pages devoted to them by Stutzer. The descriptions of the other diamond-bearing areas bring much scattered material into handy compass, though reproductions of Bauer's maps of Borneo and of Brazil without adding scales is to be regretted. The reviewer knows of no other book of this type which can compare with the present one for completeness, and it is unfortunate that the current American price of some $13.50 (allowing for the recently announced 25 per cent discount) must make ownership of the volume so difficult for most of us.

D. Jerome Fisher

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